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Key Points:

- Fault creep rates and deep slip rates are inferred from GPS data
- Creep limits slip deficit on the Bartlett Springs fault but not the Maacama
- GPS and geologic slip rate estimates agree for San Andreas but not Maacama

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Correspondence to:

J. R. Murray,
jrmurray@usgs.gov

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Slip rates and spatially variable creep on faults of the northern San Andreas system inferred through Bayesian inversion of Global Positioning System data

Jessica R. Murray¹, Sarah E. Minson², and Jerry L. Svarc¹
¹U.S. Geological Survey, Menlo Park, California, USA, ²Seismological Laboratory, Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena, California, USA

Abstract Fault creep, depending on its rate and spatial extent, is thought to reduce earthquake hazard by releasing tectonic strain aseismically. We use Bayesian inversion and a newly expanded GPS data set to infer the deep slip rates below assigned locking depths on the San Andreas, Maacama, and Bartlett Springs Faults of Northern California and, for the latter two, the spatially variable interseismic creep rate above the locking depth. We estimate deep slip rates of 21.5 ± 0.5 , 13.1 ± 0.8 , and 7.5 ± 0.7 mm/yr below 16 km, 9 km, and 13 km on the San Andreas, Maacama, and Bartlett Springs Faults, respectively. We infer that on average the Bartlett Springs fault creeps from the Earth's surface to 13 km depth, and below 5 km the creep rate approaches the deep slip rate. This implies that microseismicity may extend below the locking depth; however, we cannot rule out the presence of locked patches in the seismogenic zone that could generate moderate earthquakes. Our estimated Maacama creep rate, while comparable to the inferred deep slip rate at the Earth's surface, decreases with depth, implying a slip deficit exists. The Maacama deep slip rate estimate, 13.1 mm/yr, exceeds long-term geologic slip rate estimates, perhaps due to distributed off-fault strain or the presence of multiple active fault strands. While our creep rate estimates are relatively insensitive to choice of model locking depth, insufficient independent information regarding locking depths is a source of epistemic uncertainty that impacts deep slip rate estimates.

1. Introduction

The San Andreas Fault System in Northern California consists of three subparallel strands, the San Andreas (SAF), Maacama (MF), and Bartlett Springs (BSF) (Figure 1). While relatively less is known about the deformation rates associated with the MF and BSF than for their southeastward extensions in the San Francisco Bay Area (the Hayward and Calaveras Faults, respectively), the most recent probabilistic seismic hazard assessment for California, the Uniform California Earthquake Rupture Forecast, version 3 (UCERF3), considers them capable of nucleating and participating in earthquakes of up to approximately $M7.5$ [Field *et al.*, 2014].

A common conceptual model for interseismic deformation due to strike-slip faulting is one in which a fault is locked from the Earth's surface to some depth, the "locking" depth, and slips freely below that depth in response to relative plate motion. We refer to the slip rate below the locking depth as the deep slip rate. The seismogenic zone, located above the locking depth, accumulates a slip deficit relative to the deeper, freely slipping fault at a rate equal to the fault's deep slip rate. A moment deficit, calculated from the product of the slip deficit, the area over which that deficit occurs, and the shear modulus, accumulates over time and is expected to be recovered in future earthquakes. There is often good agreement between long-term slip rate estimates from geologic observations spanning millennia and deep slip rate estimates based on geodetic data spanning decades. UCERF3 used slip rate information inferred from both geologic and geodetic data when calculating moment deficit rates used in forecasting California earthquake probabilities [Field *et al.*, 2014].

Surface creep has been observed on several faults of the San Andreas system in Northern California, including on the MF and BSF [McFarland *et al.*, 2009; Tong *et al.*, 2013], and may extend to depth over some or all of the seismogenic zone. Creep has the potential to substantially reduce the rate at which a slip deficit accumulates or the size of the area over which a slip deficit exists. These conditions would be expected to

Figure 1. Map of the study area. Heavy black lines show model geometry surface traces; white squares are end points of segments spanned by summed Green's functions (see text for details). Red lines are mapped fault traces [U.S. Geological Survey and California Geological Survey, 2006; Lienkaemper, 2010]. Alignment array sites are labeled with site IDs [McFarland et al., 2009]; CGPS station P190 is labeled for reference (see text). SGPS: survey-mode (campaign) GPS sites; CGPS: continuous GPS sites. BSF: Bartlett Springs Fault; CF: Calaveras Fault; FI: Farallon Islands; FR: Fort Ross; GVF: Green Valley Fault; HF: Hayward Fault; LP: Lake Pillsbury; MF: Maacama Fault; PA: Point Arena; PR: Point Reyes; RCF: Rogers Creek Fault; SAF: San Andreas Fault. BSLP is located at Lake Pillsbury; MWIL is located in the town of Willits; MUKI is located in the town of Ukiah.

extend the time between major earthquakes and/or limit earthquake size, thus reducing seismic hazard posed by a fault. In the context of a creeping fault, the term “transition depth” could be used in place of “locking depth” to emphasize that the slip behavior transitions from some combination of creeping and locked to fully slipping with increasing depth. However, in this paper we opt to retain the term locking depth as it is an appropriate description of our model parameterization and is consistent with commonly used terminology. We use the term “creep rate” to refer to the rate of any interseismic slip that occurs above the locking depth.

UCERF3 accounts for the effects of fault creep using a spatially variable-scale factor that depends on the ratio of surface creep rate to long-term slip rate [Weldon et al., 2013]. This factor is applied to reduce either the slip rate or the seismogenic area used in the earthquake rate models. In order to determine the scale factors, Weldon et al. [2013] adopted simplified creep rate versus depth profiles based on studies of the Hayward Fault [Savage and Lisowski, 1993; Schmidt et al., 2005] and the SAF near Parkfield, CA [Murray et al., 2001].

These profiles, which are characterized by a decrease in creep rate with depth, were applied to all creeping faults in the UCERF3 deformation model.

Detailed observations regarding the spatial variability of creep from more faults would help validate the general applicability of this approach as well as enable more precise assessment of fault-specific seismic hazard. Kinematic models derived from geodetic data suggest that creep rates on the Hayward and Calaveras Faults can be a sizable fraction of the deep slip rate but vary substantially both along strike and with depth [Simpson *et al.*, 2001; Manaker *et al.*, 2003; Schmidt *et al.*, 2005; Evans *et al.*, 2012; Shirzaei and Bürgmann, 2013]. Whether surface creep observed on the MF and BSF extends to seismogenic depths and shows similar spatial variability has been less well studied. While Freymueller *et al.* [1999] modeled GPS data to infer that the BSF creeps at its deep slip rate at all depths, thus accumulating no moment deficit, their data set included only one pair of stations spanning that fault at distances capable of recording creep at seismogenic depths (e.g., distances $< \sim 10$ km).

The goal of the study we present here is to assess quantitatively the rates and spatial distribution of creep on the MF and BSF and interpret this information in the context of these faults' deep slip rates. This study has required two main components: data collection and deformation modeling. First, we greatly increased the coverage of GPS observations in the region by establishing a new, spatially dense campaign GPS network sensitive to creep at depth on the MF and BSF. We also reoccupied additional sites across the study region for which we could draw upon previously archived data in order to improve the quality and quantity of velocity estimates on a regional scale to constrain better the SAF, MF, and BSF deep slip rates. Second, we used a Bayesian inversion methodology to infer the spatial distribution of creep rate on the seismogenic portions of the MF and BSF along with the slip rates on the deeper extensions of the SAF, MF, and BSF. Even when the geographic coverage of geodetic observations is good, inverting the data for spatially variable fault creep is underdetermined. Least squares inversions used to infer slip distributions are typically regularized using spatial smoothing or solution damping [e.g., Harris and Segall, 1987; Du *et al.*, 1992]. The relative weighting of data fit versus solution smoothness strongly influences the resulting creep rate distributions and moment deficit rates inferred from them. In contrast, Bayesian inversion does not require regularization but instead enables us to examine the creep rate distribution at the highest spatial resolution possible given the available data [Minson *et al.*, 2013]. Furthermore, Bayesian inversion allows incorporation of arbitrary prior information, for example, regarding the sense of slip on faults of interest. By providing the posterior distribution for estimated parameters given data and prior information, the Bayesian inversion allows us to more completely evaluate the uncertainties of our rate estimates and thus whether substantial creep exists at seismogenic depths.

2. Previous Slip Rate and Creep Rate Estimates for Northern San Andreas System Faults

Fault slip rates can be estimated both from geologic and geodetic observations. Each approach has advantages and disadvantages [Parsons *et al.*, 2013]. While traditionally fault slip rate estimates used in seismic hazard assessments have been largely based on geologic data, the most recent assessment for California, UCERF3, also included deformation models based on geodetically constrained deep slip rates, although geologic slip rate estimates remain heavily weighted [Parsons *et al.*, 2013].

For faults of the San Andreas system in our study area (approximately 38.75°N–39.75°N) only a few geologic slip rate estimates are available (Figure 2; see Figure 1 for locations). An average late Pleistocene slip rate of 16–24 mm/yr was estimated for the SAF using measurements and dating of offset marine terraces near Point Arena, CA [Prentice, 1989]. This result is consistent with more recent work ~ 80 km to the south near Fort Ross where dating of an offset stream channel yielded an average slip rate of 19 ± 4 mm/yr over the past ~ 5000 years [Prentice *et al.*, 2001]. Sickler *et al.* [2005] estimated a long term slip rate of $8.7 - 13.4$ mm/yr over the past ~ 4700 years for the Maacama Fault near Ukiah based on trenching, cone penetrometer, and seismic refraction data. Based on recent trenching near Willits, C. S. Prentice *et al.* (Late Holocene slip rate and age of a prehistoric earthquake along the Maacama Fault near Willits, Mendocino County, Northern California, submitted to *Bulletin of the Seismological Society of America*, 2014; hereafter referred to as C. S. Prentice *et al.*, submitted manuscript, 2014) estimate a Maacama Fault slip rate of 6.4–8.6 mm/yr over the 560–690 years prior to 2000 C.E. McLaughlin *et al.* [2012] used offset volcanic units and magnetic anomalies to estimate a

Figure 2. Comparison of geologic slip rate and geodetic deep slip rate estimates. Geologic slip rates are given as ranges stated in corresponding publications; 1: Prentice [1989], 2: Prentice et al. [2001], 3: Sickler et al. [2005], and 4: C. S. Prentice et al. (submitted manuscript, 2014). UCERF3 weighted average was calculated using the values from the geodetic and geologic deformation models for SAF, MF, and BSF segments in our study area. The weighted average was computed using the logic tree weighting assigned to each deformation model [Parsons et al., 2013].

long-term slip rate range of ~ 5.5 – 8.3 mm/yr over the past ~ 3.17 Ma for the MF outside of our study area to the south. No geologic slip rate estimates are available for the BSF. Preferred geologic slip rates assigned to the MF (9 mm/yr) and BSF (4 mm/yr) in the UCERF3 deformation models are based on extrapolation of rates from adjacent fault segments located to the southeast; the SAF geologic slip rate used in UCERF3 was 24 mm/yr based on the studies described above [Dawson, 2013].

Common approaches for estimating strike-slip fault slip rates from geodetic data include 2-D models that assume infinitely long faults that slip at the long-term rate below the locking depth [Savage and Burford, 1973], 3-D models that approximate long-term motion using quasi-infinite dislocations to represent the deep slipping fault [Manaker et al., 2003; Schmidt et al., 2005], and models in which relative motion due to rotation of fault-bounded blocks is used to fit geodetically recorded elastic strain via a back slip formulation with assumed or estimated locking depths [McCaffrey, 2005; Meade and Hager, 2005]. The following is a

summary of existing geodetic slip rate estimates for the northern San Andreas system; in all cases the Earth's crust was assumed to behave as a homogeneous elastic half-space. Figure 2 shows these slip rate estimates in comparison to those from geologic data.

The data set used by Freymueller et al. [1999] is most similar to ours in terms of geographic extent. They simultaneously optimized the locking depths of the SAF, MF, and BSF and estimated the deep slip rates of each fault to fit GPS observations using 2-D models. Freymueller et al.'s [1999] preferred deep slip rates were 17.4 mm/yr, 13.9 mm/yr, 8.2 mm/yr, and locking depths were 14.9 km, 13.4 km, and 0 km for the SAF, MF, and BSF, respectively. Note that this result implies that the BSF in their model slips freely at all depths. Prescott et al. [2001] also used a 2-D approach to model the slip rates on the southeastward extensions of these faults in the northern San Francisco Bay Area. Their preferred model held the locking depths fixed at the values used in the Working Group on California Earthquake Probabilities [1999] study (11 km, 12 km, and 14 km for the SAF, Rogers Creek Fault, and Green Valley Fault, respectively; see Figure 1 for locations), and the corresponding deep slip rate estimates were 20.8, 10.3, and 8.2 mm/yr. As discussed by Freymueller et al. [1999] and Prescott et al. [2001], there are trade-offs between deep slip rate and locking depth on a given fault as well as among these parameters for multiple closely spaced, subparallel, strike-slip faults. The integrated slip rate across the three faults of approximately 40 mm/yr found by these two studies is better constrained than the individual faults' deep slip rates. This total rate is consistent with geodetically constrained estimates of Pacific-Sierra Nevada relative block motion within our study region of 38–40 mm/yr [Argus and Gordon, 2001; Sella et al., 2002; d'Alessio et al., 2005; Titus et al., 2011].

McCaffrey [2005] developed a California state-wide block model constrained by GPS data as well as geologic slip rate estimates, slip vectors derived from moment tensor solutions, spreading rates, and transform azimuths. In the vicinity of the northern San Andreas system the GPS data set primarily consists of velocities

Figure 3. Comparison of creep rate estimates from alignment array data [McFarland *et al.*, 2009] and results from Bayesian inversion of GPS data (this study) at corresponding locations. Alignment array site names correspond to Figure 1. Error bars are 2σ . Diagonal line indicates perfect agreement between two sets of estimates.

published by Freymueller *et al.* [1999]. While a nominal locking depth of 20 km was assumed for all faults, the degree of fault coupling was estimated for each fault in the inversion such that the effective locking depth for a given fault could range from 1 km (creeping) to 20 km (fully locked). Deep slip rate estimates for the SAF, MF, and BSF were 21 mm/yr, 7.4 mm/yr, and 7.3 mm/yr, respectively, with a SAF effective locking depth of ~ 12 km and the MF and BSF creeping below ~ 1 km. Several deformation models that incorporated GPS data were developed for UCERF3 [Parsons *et al.*, 2013]. Some used an elastic block approach, while others modeled the deformation as occurring in a faulted continuum, but in all cases geologic and geodetic data were fit simultaneously with more weight given to fitting the geologic slip rates where well-constrained geologic estimates exist. The approximate range of estimated deep slip rates for the SAF, MF, and BSF were 16–22.5 mm/yr, 9.4–12 mm/yr, and 3.8–7.5 mm/yr, respectively.

Alignment array, Interferometric Synthetic Aperture Radar (InSAR), and geologic observations indicate that both the MF and BSF creep at the surface and possibly at depth. Alignment array locations in our study area are shown in Figure 1, and rates estimated from these data are summarized in Figure 3. Alignment array measurements on the MF have recorded a surface creep rate of 4.3 mm/yr since 1993 at Ukiah, and 5.7 mm/yr since 1991 at Willits [McFarland *et al.*, 2009]. Subsequent to the inference of Freymueller *et al.* [1999] that the BSF might be creeping at all depths, an alignment array was established at Lake Pillsbury on the BSF in 2005. An average creep rate of 3.1 mm/yr has been recorded at this site; the creep occurs episodically, similar to what has been observed on the Green Valley Fault to the southeast [McFarland *et al.*, 2009]. Alignment arrays established more recently to the north and south on the BSF have yet to record significant creep [McFarland *et al.*, 2009]. Tong *et al.* [2013] used Advanced Land Observing Satellite, Phased Array type L-band Synthetic Aperture Radar (ALOS PALSAR) interferograms to estimate line-of-sight velocities throughout the San Andreas system and used these to compute horizontal offset rates across creeping faults. To first order, their results are consistent with alignment array measurements on the MF and BSF; however, at several locations along these faults they infer apparent left-lateral creep. They note that this may be due to vertical motion and that in most cases this motion is not significant within errors.

Other observations also suggest that the Bartlett Springs Fault zone near Lake Pillsbury is capable of creep. For instance, serpentinite observed in a fault exposure near Lake Pillsbury shows similar mineralogy to that of cuttings from the portion of the San Andreas Fault Observatory at Depth hole that passed through a creeping strand of the SAF [Moore *et al.*, 2011]. In addition, at least two of the creep events near Lake Pillsbury since 2005 have been accompanied by fresh en echelon cracks in pavement near the instrument site (F. S. McFarland, personal communication, 2013).

3. Data

While estimating the deep slip rate from geodetic data requires broadly spaced measurement sites spanning several locking depths (typically 10–15 km for faults of the San Andreas system) on either side of the fault, densely spaced observations within one locking depth of the fault trace are needed to infer the creep rate distribution in the seismogenic zone. Prior to this study the primary geodetic data set for the MF/BSF

Figure 4. Observed GPS velocities. Error ellipses are 95% confidence. Black lines are mapped faults [U.S. Geological Survey and California Geological Survey, 2006; Lienkaemper, 2010]. (a) Velocities in North America fixed reference frame. Map projection is about the Pacific-North America pole of relative rotation estimated by *d'Alessio et al.* [2005]. (b) Velocities in a Sierra Nevada/Great Valley fixed reference frame. Map projection is about the Pacific-Sierra Nevada/Great Valley pole of relative rotation estimated by *d'Alessio et al.* [2005]. (c) Velocities parallel (top) and perpendicular (bottom) to direction of Pacific-Sierra Nevada relative motion estimated using Euler pole calculated by *d'Alessio et al.* [2005]. Positive velocities indicate northwestward and northeastward motion in top and bottom panels, respectively. Heavy black lines indicate distances on x axis spanned by mapped surface traces of the three major faults.

region was that used by *Frey Mueller et al.* [1999] which consisted of two profiles of stations roughly perpendicular to the three faults of the northern San Andreas system, one extending northeastward from Point Arena through Ukiah and the other about 15 km to the northwest passing through Willits and Lake Pillsbury. These profiles each included 2–3 sites on either side of the MF within ~10 km of the surface trace. However, this data set included only one fault-crossing pair of sites within ~10 km of the BSF. The continuous GPS (CGPS) sites of the Plate Boundary Observatory (PBO) which came online in the mid-2000s and a small number of sites from the Bay Area Regional Deformation (BARD) network which date back to the mid-1990s provide good broad-scale coverage in the region but are insufficient for measuring fault creep.

To address this need, in 2005 the U.S. Geological Survey (USGS) Earthquake Science Center established a spatially dense campaign, or survey-mode, GPS (SGPS) network spanning the MF and BSF consisting of 99 new sites. In addition, we have leveraged older data in publically available archives by resurveying 34 SGPS sites that span our geographic region of interest in order to extend their time series. Of these sites, a total of 105 met our data quality criteria described in the supporting information. Our full data set consists of velocities from these sites as well as 34 CGPS sites of the PBO and BARD networks. Given the scarcity of GPS sites to the west of the SAF, we include two CGPS stations to the south of our study area, one at Point Reyes and one on the Farallon Islands, to provide additional constraint on the SAF slip rate. Our data set also includes CGPS sites extending east into the Great Valley (Figure 1).

We processed the GPS data using the GIPSY-OASIS II software developed by NASA's Jet Propulsion Laboratory. We applied regional filtering to reduce common mode scatter [Wdowinski *et al.*, 1997; Dong *et al.*, 2006] and then

Figure 4. (continued)

applied the time series analysis methodology of *Langbein* [2004] to estimate average velocities for each GPS site and velocity uncertainties that account for temporally correlated noise. Details of the data processing and time series analysis are provided in the supporting information. The resulting velocity field used in the modeling is shown in Figure 4a and is provided in Tables S2 and S3 in the supporting information.

As an initial step we consider the GPS velocity field in the context of the regional tectonic setting. Figure 4a shows the North America-fixed velocity field and regional faults in an oblique Mercator projection about the North America-Pacific pole of relative motion calculated by *d'Alessio et al.* [2005]. Figure 4b shows the velocities in a Sierra Nevada-fixed reference frame, plotted using an oblique Mercator projection about the Sierra Nevada-Pacific pole [*d'Alessio et al.*, 2005]. It is evident that, in general, the three major faults are neither parallel to each other nor to the motion predicted from either pole of rotation. Although there is little net contraction between the Great Valley and the coast (Figure 4b), the GPS velocities suggest a component of extension west of the BSF transitioning to compression west of the MF (Figure 4c, bottom). As discussed by *d'Alessio et al.* [2005], however, the degree of fault normal convergence depends not on the orientation of velocity vectors compared to the direction of relative plate motion but on the velocity orientation relative to the local fault strike which varies throughout our study area and may reflect reactivation of preexisting structures [*McLaughlin et al.*, 2012; *Thomas et al.*, 2013]. Using infinitely long 2-D screw dislocations to model regional deformation [e.g., *Freymueller et al.*, 1999] implicitly assumes that the regional faults are parallel and predicts GPS velocities that are fault parallel. Given that neither of these characteristics applies to the northern San Andreas system, we choose instead to use a dislocation model that accommodates changes in fault strike and the full three-dimensional velocity field.

4. Modeling

4.1. Model Geometry

4.1.1. Representation of Fault Using Dislocations

We use the GPS velocities to estimate the SAF, MF, and BSF deep slip rates as well as spatially variable creep rates on the MF and BSF above the locking depth. To do so we represent the faults using rectangular dislocations in a homogeneous elastic half-space (assuming a Poisson's ratio of 0.25) and use the expressions from *Okada* [1985] to relate the motion at the GPS sites to fault slip.

Although geodetic data can be used in a nonlinear inversion to estimate source dimensions, orientation, and location, here we draw upon independent information to specify some of these parameters. Our station

Figure 5. (a) Example dislocation geometry used in Bayesian model class selection and synthetic tests (see text and Text S1 in the supporting information for details). Light gray subfaults are allowed to creep; dark gray represents a small portion of the quasi-infinite dislocations used to represent deep slip on the three faults. In the Bayesian model class selection the locking depths and, for the MF and BSF, the dip of the creeping zones were allowed to vary. (b) Dislocation geometry used in Bayesian inversion for deep slip rates and spatially variable creep rates. Heavy black lines outline groups of subfaults for which a uniform creep rate was estimated by summing their Green's functions. LD: locking depth.

distribution determines the length of the fault segments used in the models, while mapped surface traces fix the location and strike of the faults (Figure 1). In the study area we adopt the simplified SAF trace used in the UCERF3 deformation models which was derived from the Southern California Earthquake Center Community Fault Model [Dawson, 2013]. For the MF and BSF we use model surface traces developed by R. Simpson (personal communication, 2012) based on information from the *U.S. Geological Survey and California Geological Survey* [2006] and on detailed geologic mapping [Lienkaemper, 2010], respectively.

For the MF and BSF we model creep in the seismogenic zone using grids of rectangular dislocations (subfaults) that extend downward from each fault's surface trace to its locking depth at a specified dip (see, for example, Figure 5a). The choice of locking depths and dips are discussed below. The number of subfaults along strike was specified by the number of segments required to approximate the mapped traces; the majority of subfaults are 5 km in along-strike dimension. The MF and BSF each have four rows of subfaults in the downdip direction. The absence of creep on the SAF in our study area is supported both by alignment array measurements [McFarland *et al.*, 2009] and the lack of microseismicity which is typically abundant on creeping faults [Rubin *et al.*, 1999]. Given this, along with the

sparse spatial coverage of GPS sites near the SAF in our data set, we do not include creeping dislocations for this fault in the model.

The deep slip rate on each of the three faults is modeled using large vertical rectangular dislocations that extend downward from the fault's locking depth. For the MF and BSF the horizontal locations of the upper edges of their deep dislocations therefore coincide with the deepest edge of their creeping dislocations. The SAF is assumed to be vertical, so its deep dislocations are directly beneath the surface trace. Multiple deep dislocations are used for each fault in order to allow for variations in strike within the study area that affect the location of faults relative to GPS stations; outside of this region the deep dislocations extend northwest and southeast with a uniform strike of N32°W. The deep dislocations for each fault have downdip and cumulative along-strike dimensions of ~50,000 km in order to approximate the signal of an infinitely long strike-slip fault without introducing edge effects. The tectonic history of the San Andreas

system, in particular, the passage of the Mendocino Triple Junction and resulting slab window [Dickinson and Snyder, 1979], has shaped the region's complex structural fabric (see, for example, Furlong and Schwartz [2004], McLaughlin *et al.* [2012], and references therein). The fault parameterization we employ here is a simplified representation of the northern San Andreas system which we feel captures first-order features of the regional deformation and is in keeping with results from seismic reflection and refraction studies that suggest the SAF and MF cut through the entire crust [Henstock *et al.*, 1997; Hole *et al.*, 2000]. We discuss this further in section 6.1.2.

For each dislocation in our model geometry, we calculate Green's functions relating GPS velocities to creep rate or deep slip rate on that dislocation [Okada, 1985]. We wish to estimate a single uniform deep slip rate for each of the SAF, MF, and BSF. Under the elastic assumption, the surface motion is linearly related to slip at depth. Therefore, for each GPS velocity we can sum the Green's functions for the deep dislocations corresponding to a given fault and use the summed Green's functions to fit the velocities with a single deep slip rate value for that fault. As discussed in section 4.2 we use the same approach to reduce the number of creep rate parameters for the MF and BSF while still allowing the model geometry to follow the strike prescribed by the surface traces. We do not apply any constraint to the cumulative deep slip rate across the three faults.

4.1.2. Assigning Dips and Locking Depths

To fully specify the model fault geometry, we must assign locking depths to each of the three faults as well as specify the dip of the creeping subfaults for the MF and BSF. These quantities can be optimized through a nonlinear inversion while simultaneously estimating deep slip rates and the creep rate distribution; however, strong trade-offs among these parameters lead to poorly constrained solutions. Instead, we look to independent information from seismicity to gain insight into fault locking depth and fault dip above that depth.

Waldhauser and Schaff [2008] relocated seismicity for all of Northern California using waveform cross-correlation and double-difference techniques. The resulting relative relocations for the MF and BSF are characterized by diffuse seismicity with discontinuous structures that might be interpreted as fault strands of limited spatial extent. Some of these structures are nearly vertical and others dip to the northeast. Clear through-going MF or BSF surfaces are difficult to identify. Within 10 km horizontally of the MF and the BSF surface traces in our study area, 95% of Waldhauser and Schaff's relocated seismicity is shallower than 9.2 km and 12.9 km depth, respectively. Following the approach of d'Alessio *et al.* [2005], we use these values rounded to the nearest integer to fix the locking depth in our model geometry to 9 km for the MF and 13 km for the BSF. Note that for the MF we only considered seismicity north of 38.9°N to avoid earthquakes associated with the Geysers geothermal field which would have biased the depth estimate to shallow values.

While the relocated seismicity gives an indication of the locking depth, its diffuse nature limits its usefulness for inferring the dip of the MF and BSF in the seismogenic zone. Furthermore, the very limited number of earthquakes in the vicinity of the SAF precludes using seismicity as a guide to locking depth there. To address this, we used the GPS data to infer the SAF locking depth and the dips of the MF and BSF by employing Bayesian model class selection following the approach developed by Minson *et al.* [2014]. Based on the results of this analysis, which is described in detail Text S1 in the supporting information, we assign a locking depth of 16 km for the SAF and dip of 60°NE and 90° for the creeping portions of the MF and BSF, respectively.

4.2. Bayesian Inversion and Model Resolution

Having selected geometries for the three faults, we employ Bayesian inversion [Jaynes, 2003; Tarantola, 2005; Beck, 2010; see also discussion in Minson *et al.*, 2013] to estimate deep slip rates and spatially variable creep rates. This approach has several advantages over traditional regularized least squares approaches [e.g., Menke, 2012] to estimating fault slip distributions. Rather than finding only the best fitting solution, we sample the full solution space to obtain the posterior probability density function (PDF) that describes the relative plausibility of all possible deep slip rate and creep rate values conditional on prior constraints and observations. The resulting PDF enables us to evaluate the uncertainties on model parameters. This approach does not require regularization and therefore avoids the need to choose subjectively weighted constraints such as spatial smoothing that may lead to spurious slip estimates or inaccurate inferences of spatial variability in model parameters. At the same time, any sort of prior information can be included to incorporate the knowledge we do have regarding the physics of the problem. A drawback to Bayesian

methods is their computational expense which arises because the number of samples (i.e., evaluations of the forward model) required to characterize the posterior distribution of the solution space increases nonlinearly with the number of model parameters [Bellman, 1957]. Here we use the Cascading Adaptive Transitional Metropolis in Parallel (CATMIP) algorithm [Minson *et al.*, 2013] which was designed to handle the large number of model parameters inherent in estimating spatially variable slip (or in our case, spatially variable creep rate). CATMIP combines elements of simulated annealing and genetic algorithms with the Metropolis algorithm [Metropolis *et al.*, 1953] into a parallel sampling algorithm in order to compute the inverse.

Taking this approach, we estimate the deep slip rates on the SAF, MF, and BSF as well as the creep rates on the seismogenic zone of the MF and BSF assuming the locations, orientations, and locking depths described above. As part of the Bayesian analysis, we include a linear regression to estimate a three parameter translation that aligns the data and model reference frames. Following the algorithm of Minson *et al.* [2013], we solve for both strike-slip and dip-slip rates. The SAF, MF, and BSF are assumed to be predominantly strike-slip faults, so we expect any dip-slip component to be small and assign these parameters a Gaussian prior distribution with mean zero and standard deviation of 1 mm/yr. For the strike-slip component we use a uniform prior that is truncated on its lower end at -1 mm/yr (where the negative indicates left-lateral slip).

Subdividing the MF and BSF above their locking depths into grids of subfaults (Figure 5) enables us to estimate spatially variable creep rate. The appropriate subfault sizes depend on the model resolution [e.g., Page *et al.*, 2009; Barnhart and Lohman, 2010]. We evaluated the implications of model resolution for assigning subfault size as a function of location on the MF and BSF in two ways: by inspection of the posterior probabilities for creep rate estimates from inversions using real data and through synthetic tests.

When we carry out the Bayesian inversion using the real data and a model geometry that has relatively small subfaults (approximately 5 km along strike and ranging from 2 to 5 km in downdip dimension), the marginal posterior distributions for strike-slip creep rate on most subfaults are highly peaked at zero, approximately following a Dirichlet distribution. This does not mean that these subfaults are not creeping. Rather, as discussed in Minson *et al.* [2013], inversions utilizing a back slip constraint will have a Dirichlet-shaped posterior distribution when creep rates on multiple-subfault trade-off with each other due to poor spatial resolution. While the motion on each subfault is poorly constrained, the total creep across multiple subfaults is well constrained. We find that increasing subfault size results in more nearly Gaussian marginal posterior distributions, demonstrating that the data can constrain the total creep rate over larger portions of the fault. Therefore, although Bayesian inversion does not require spatial smoothing constraints in the way that regularized least squares inversion does, the need to address limited spatial resolution in some way is unavoidable. Here we describe the approach we have taken.

To evaluate the resolving power of our station distribution, we carried out synthetic tests in which data were generated given a hypothetical distribution of deep slip rates and creep rates using a grid of small subfaults similar to that described in section 4.1 (e.g., Figure 5a). Normally distributed random noise based on the observed data covariance was added to the synthetic data. We conducted a series of tests in which the Green's functions for groupings of neighboring subfaults were summed such that a uniform creep rate was estimated for each grouping. For each inversion with summed Green's functions, we compared the estimated deep slip rates to their true values and each estimated creep rate to the average of the true creep rates over the area spanned by the corresponding Green's function.

When Green's functions are summed over small numbers of subfaults, thus effectively spanning a small portion of the fault's surface area, the resulting posterior distributions for creep rate have large standard deviations and are thus not very informative. While the distribution may encompass the true creep rate value at 95% confidence, the mean or mode of the distribution may not be a good representation of the true value. We adopted the minimum grouping of subfaults that produced creep rate estimates that matched the true synthetic rates and were well constrained in the sense of having posterior distributions that did not indicate substantial trade-off with neighboring subfaults.

We found that in order to obtain well-constrained creep rate estimates that matched the true rates; we needed to sum Green's functions both along strike and downdip. The resulting grouping of subfaults for summed Green's functions is shown schematically by the heavy outlines in Figure 5b. In the downdip direction we could retain only two depth increments, from the Earth's surface to 5 km and from 5 km to the locking depth. Under this parameterization, each creep estimate represents depth-averaged behavior over

Figure 6. Mode of posterior distribution for creep rates and deep slip rates obtained by Bayesian inversion. For clarity, only a portion of each deep dislocation is shown. (a) Strike slip and (b) dip slip. Note different color scales.

the corresponding increment. The number of subfaults over which to sum Green's functions along strike correlated well with the near-fault station distribution and permits estimation of along-strike creep rate variation in the upper 5 km. However, attempting to allow along-strike variation from 5 km depth to the locking depth resulted in substantial trade-offs among creep rate estimates on neighboring subfaults. Therefore, we have elected to estimate a single uniform creep rate for each of the MF and BSF from 5 km depth to the faults' respective locking depths. Using summed Green's functions, the posterior distributions for most creep rate estimates are approximately Gaussian rather than being peaked at zero. For consistency in model parameterization, we sum Green's functions identically for strike-slip and dip-slip components. The parameters defining a simplified set of subfaults that describe this geometry (fine outlines in Figure 5b) are given in the dislocation format of Okada [1985] in Table S6 in the supporting information; Table S7 includes information for matching these subfaults to summed Green's functions.

Using small subfaults and applying a spatial average to each creep rate distribution in the resulting posterior PDF could be used as an alternative to summing Green's functions [Duputel et al., 2014]. Both spatially filtering the slip distribution a posteriori and spatially varying the fault discretization by summing Green's function require identifying appropriate averaging distances in the two fault-plane dimensions. However, summing the Green's functions has the added benefit of reducing the number of model parameters and thus the computational expense of the inversion.

Our model now consists of 30 parameters, 15 for strike-slip and 15 for dip-slip rates. For each slip component there are six near-surface and one intermediate depth creep rate estimates for the MF, four near-surface and one intermediate depth creep rate estimates for the BSF, and one deep slip rate each for the SAF, MF, and BSF. Following the guidelines suggested by Minson et al. [2013] we use 10,000 samples and a Markov chain length of 5,000 in the Bayesian inversion. Markov chain length refers to the number of random walk steps in each chain. For larger numbers of model parameters, more samples and longer chains are required to adequately sample the solution space [Minson et al., 2013].

5. Results

Figure 6 shows the mode of the deep slip rate and creep rate posterior PDFs obtained through the Bayesian inversion. Histograms depicting the posterior distribution for creep rates are shown in Figure 7. The mode is the most frequent value of a probability distribution and for a Gaussian distribution would be the same as the mean. The majority of the rate estimates have posterior distributions that look roughly Gaussian. However, for the rate estimates that have posterior distributions peaked near zero, the mode captures the fact that from the ensemble of possible rate solutions the estimate for this rate parameter was usually zero.

Figure 7. Posterior distributions for creep rate estimates (mm/yr). X axis labels give Green's function index and correspond to numbered areas on the fault diagrams. (a) Strike slip and (b) dip slip. Dashed lines outline portions of deep slip dislocations.

Figure 8. Posterior distributions for deep slip rate estimates (mm/yr) and scatterplots showing trade-offs among the deep slip rates for the three faults. (a) Strike slip and (b) dip slip. Note different scales.

The deep slip rate estimates and two-sigma uncertainties for the SAF, MF, and BSF are 21.5 ± 0.5 , 13.1 ± 0.8 , and 7.5 ± 0.7 mm/yr below 16 km, 9 km, and 13 km depth, respectively. While the uncertainties on these estimates are small, there are strong trade-offs among the rate estimates for the three faults (Figure 8). The sum of the deep slip rates across the region is 42.1 ± 0.3 mm/yr.

Creep rate estimates for both the MF and BSF show along-strike variation with locally high values, especially at the NW end of the two faults. The estimated uniform creep rate from 5–9 km depth on the MF, 2.9 mm/yr, is less than the rates estimated for much of the shallower portion of this fault. The opposite is true for the

Figure 9. GPS velocity residuals (observed minus predicted) for model shown in Figure 6. Scale vectors (red and blue) indicate 3 mm/yr; ellipses are 95% confidence. Red vectors indicate velocities not fit at 2σ . Heavy lines show model surface traces color coded by estimated creep rate in upper 5 km. Color scale is the same as Figure 6a. Red lines are mapped fault traces [U.S. Geological Survey and California Geological Survey, 2006; Lienkaemper, 2010].

BSF whose estimated uniform creep rate of 7.7 mm/yr from 5 to 13 km depth exceeds the estimated creep rates for most of the fault above 5 km and is similar to this fault's estimated deep slip rate.

For comparison, we carried out a regularized least squares inversion in which the MF and BSF creeping portions were discretized into smaller subfaults, and spatial smoothing was applied. Like the results of the Bayesian inversion, the creep rate distributions inferred from this analysis (Figure S1 in the supporting information) exhibit clear along-strike variations in magnitude; in the smoothed inversion these variations extend to the locking depth. A least squares inversion using the coarse dislocation parameterization of the Bayesian inversion, no spatial smoothing, and a lower bound of 0 mm/yr on creep rates and slip rates to enforce right-lateral slip, produced nearly identical slip rate and creep rate estimates to those of the Bayesian inversion. However, we favor the Bayesian approach because it permits rigorous inclusion of prior information, for example, that the sense of slip is right lateral, and more fully characterizes the uncertainties on estimated parameters.

While the estimated dip-slip creep rates above 5 km depth suggest a northwestward transition from reverse creep to normal creep on the MF and from normal creep to reverse creep on the BSF, most of these creep estimates are not significantly different from zero. See Table S7 in the supporting information for deep slip rate and creep rate estimates.

The fit to the data using the results shown in Figure 6 is shown in Figure 9; the predicted data are provided in Tables S2 and S3 in the supporting information. The majority of the SGPS velocities are fit at 2 sigma. The CGPS uncertainties are substantially smaller than for the SGPS sites, and a larger percentage of the CGPS stations' velocities are not fit at 2 sigma than is the case for the SGPS data. However, in most cases the CGPS residuals are small; 95% of the CGPS horizontal residual magnitudes are <1.8 mm/yr with a median value of 0.94 mm/yr.

6. Discussion

6.1. Deep Slip Rates

6.1.1. Comparison With Results of Other Studies

We first consider our deep slip rate estimates in the context of results from other studies (Figure 2). The deep slip rate of 21.5 ± 0.5 mm/yr (all stated uncertainties for our results are 2 sigma) that we estimated for

the SAF is consistent with geologic slip rate estimates [Prentice, 1989; Prentice *et al.*, 2001]. It exceeds Freymueller *et al.*'s [1999] estimate of 17.4 mm/yr based on GPS data, although not at 95% confidence, and is similar to deep slip rate estimates of McCaffrey [2005] and the UCERF3 deformation models [Parsons *et al.*, 2013] which used both geologic and geodetic data.

Our MF deep slip rate estimate of 13.1 ± 0.8 mm/yr exceeds the long-term slip rate of ~ 5.5 – 8.3 mm/yr estimated by McLaughlin *et al.* [2012] for the past ~ 3.17 Ma to the south of our study area. McLaughlin *et al.* [2012] note that differences between long-term rate estimates and those determined from recent data may reflect actual slip rate variations over time rather than inaccuracies in particular slip rate studies. Our MF deep slip rate estimate exceeds the geologic slip rate of 6.4–8.6 mm/yr estimated by C. S. Prentice *et al.* (submitted manuscript, 2014) at Willits but falls within the 8.7–13.4 mm/yr estimated by Sickler *et al.* [2005] near Ukiah. C. S. Prentice *et al.* (submitted manuscript, 2014) note a subsidiary fault, termed the East Willits Fault, that is located ~ 10 km northeast of the main MF trace near Willits and has geomorphic expression indicative of late Quaternary movement. Background seismicity in the Willits region is concentrated beneath this strand rather than beneath the main trace. C. S. Prentice *et al.*'s study area did not span the East Willits Fault, so their geologic slip rate estimates do not account for any movement that may be recorded on that strand. Our MF deep slip rate estimate is similar to that of Freymueller *et al.* [1999] but exceeds those based on a combination of geologic and geodetic data [McCaffrey, 2005; Parsons *et al.*, 2013]. MF geodetic slip rate estimates would be expected to exceed geologic estimates if the former encompass the contributions of multiple active fault strands.

There is no geologic slip rate estimate available for the BSF; our deep slip rate estimate of 7.5 ± 0.7 mm/yr is compatible with the UCERF3 geologic slip rate bounds of 1–9 mm/yr for the BSF which were based on northwestward extrapolation of the geologic slip rate estimates for the Concord and Green Valley Faults [Dawson, 2013]. It is also consistent with that inferred by Freymueller *et al.* [1999], McCaffrey [2005], and the UCERF3 deformation models [Parsons *et al.*, 2013]. Near-fault pairs of GPS sites, along with the Lake Pillsbury alignment array [McFarland *et al.*, 2009], show evidence for along-strike variation in surface creep rate on the BSF (Figure 10b), locally reaching ~ 8 mm/yr north of Lake Pillsbury as recorded by the fault-spanning pair of GPS sites ISLE and WOTI. This creep rate lends support to the possibility that the BSF deep slip rate is higher than the UCERF3 preferred geologic rate of 4 mm/yr [Dawson, 2013] since it is unlikely that the surface creep rate, if constant, would exceed the deep slip rate. Creep rates might temporarily exceed the deep slip rate during a creep event, and such events are observed as steps in the alignment array data from Lake Pillsbury. However, the temporal sampling of the campaign GPS at ISLE and WOTI is insufficient to determine if transient creep accelerations have occurred here during the study period.

As discussed by Freymueller *et al.* [1999] and as evident in Figure 8, there are substantial trade-offs among deep slip rate estimates for closely spaced subparallel strike-slip faults. The sum of our deep slip rate estimates for the SAF, MF, and BSF of 42.1 ± 0.3 mm/yr is expected to be better constrained than the individual slip rate estimates. This summed deep slip rate agrees within uncertainties with the estimated rate of relative motion between the Pacific plate and the Sierra Nevada/Great Valley block which ranges from 30.2 to 49.8 mm/yr (2 sigma) at the Farallon Islands [Argus and Gordon, 2001; Sella *et al.*, 2002; d'Alessio *et al.*, 2005; Titus *et al.*, 2011].

6.1.2. Epistemic Uncertainty in Deep Slip Rate Estimates

In addition to the trade-offs among deep slip rates on the three faults, there are significant trade-offs between a fault's assumed locking depth and its deep slip rate estimate; deeper locking depths are associated with higher deep slip rate estimates. Because the geodetic data are not able to independently constrain both the locking depths and deep slip rates for the three faults in our study area, we relied on inferences from seismicity to fix most of the fault geometry parameters. The deep slip rate uncertainties we obtain from the posterior PDF do not account for the epistemic uncertainty caused by unknown locking depth.

For comparison to the results obtained when the MF and BSF locking depths were determined using relocated seismicity, we repeated the Bayesian model class selection procedure described in section 4.1.2 and the supporting information using a grid search over five parameters: the locking depths for the three faults and the dips of the MF and BSF above their locking depths; the details of this analysis are described in Text S1. The best fitting locking depths for the resulting model are 14 km, 16 km, and 4 km with corresponding deep slip rates of 19.0, 16.6, and 6.0 mm/yr for the SAF, MF, and BSF, respectively (Table S5).

Figure 10

As with the results we obtained from the Bayesian inversion using fixed locking depths for the MF and BSF (Figure 6), the results shown in Table S5 suggest substantial creep on the BSF at all depths. Freymueller *et al.* [1999] came to a similar conclusion. Another way of interpreting this result is that the BSF has a shallow locking depth, around 5 km. A geodetically inferred locking depth that is substantially shallower than the depth extent of seismicity has been observed on the San Jacinto Fault in Southern California where it has been interpreted as evidence for deep creep on a portion of the fault above the brittle/ductile transition [Wdowinski, 2009].

Another feature of the results shown in Table S5 is that, in the absence of constraints on the three faults' locking depths from seismicity, the geodetic data prefer a deeper locking depth for the MF than for the SAF. Again, Freymueller *et al.* [1999] noted similar behavior. The range of models they inferred included MF locking depths as deep as 41.5 km, in which case most slip across the region was localized on this fault. They attributed this result to the smooth gradient in fault-parallel velocity from the coast to the western edge of the Great Valley (e.g., Figure 4c, top) which, while exhibiting steps at the MF and BSF due to near-surface creep, does not exhibit pronounced broader-scale inflections that would be characteristic of strike-slip faults with shallower locking depths than those we infer for the SAF and MF.

One interpretation might be that the smoothness of the observed velocity gradient reflects localization of deformation on a deep plate boundary shear zone. Alternatively, distributed shear between the SAF, MF, and BSF might obscure the locking signal from these faults in the velocity field. Castillo and Ellsworth [1993] noted the presence of earthquakes with strike-slip mechanisms between the MF and SAF north of Point Arena as a possible indication of shear strain being accommodated on additional fault strands, and Langenheim *et al.* [2013] present aeromagnetic evidence for a complex fabric of fault and fold structures between the MF and BSF. Thomas *et al.* [2013] interpret two swarms of predominantly strike-slip seismicity on a linear structure between and subparallel to the MF and BSF as evidence for initiation of a new fault in this region, perhaps by reactivation of existing structures related to earlier subduction. The existence of distributed shear across our study area, which would be mapped to fault slip in our model, would bias the deep slip rate estimates to higher values and is a possible explanation for inferred deep slip rates that exceed the geologic estimates.

Furlong *et al.* [2003] suggest that the northwestward migration of the Mendocino Triple Junction first causes crustal thickening and subsequently thinning of the crust in its wake. The bulk strain caused by such crustal thinning south of the triple junction would be recorded in the geodetic velocity field and could impact slip rates estimated from these data. Inspection of GPS velocities across our study area does not reveal clear evidence for a crustal thinning signal, as might be suggested by gradients in fault-parallel velocity aligned with the direction of migration of the triple junction. However, future work could employ more complex deformation models to test the possible influence of bulk strain or distributed off-fault shear.

Obtaining GPS velocities at more locations and with lower uncertainties would help to better characterize the patterns of strain localization associated with the SAF, MF, and BSF, although the lack of off-shore observations limits the ability to narrow the range of likely SAF locking depths and deep slip rates. Recent advances in seafloor geodetic techniques [Bürgmann and Chadwell, 2014], if implemented with long (e.g., decadal) observation histories, have the potential to provide geodetic velocity estimates for off-shore sites with sufficient precision to better constrain the rates of deformation across the northern San Andreas system and on the SAF in particular.

Another source of epistemic uncertainty is Green's function error due to assuming a uniform elastic half space. Rigorous methods for incorporating model prediction error due to inaccurate Green's functions into the Bayesian inversion framework are being explored [e.g., Minson *et al.*, 2013; Duputel *et al.*, 2014]. These efforts may, in the future, lead to approaches for incorporating epistemic uncertainty more generally and enable analyses such as that presented here to provide a more representative assessment of uncertainty on deep slip rates and/or locking depths.

Figure 10. GPS velocities parallel to fault segments spanned by each Green's function for surface-breaking subfaults. Positive velocities indicate northwestward motion. In each panel only the GPS sites within 20 km of the fault trace and whose locations project onto the segment of the fault spanned by the corresponding Green's function are shown. Note that steps at the fault are evident in most panels indicating the presence of fault creep. (a) MF and (b) BSF. Refer to Figure 7a for diagram matching Green's function indices to locations on faults. Error bars are 2σ .

Figure 11. Trade-off of strike-slip creep rate estimate for 5 km locking depth with deep strike-slip rate and strike-slip creep rate estimates above 5 km. Diagrams show portion of fault corresponding to each Green's function index. (a) MF and (b) BSF. Dashed lines outline portions of deep slip dislocations.

6.2. Creep Rates

6.2.1. Features of the Inferred Creep Model

The results of the Bayesian inversion suggest that both the MF and BSF exhibit significant creep in the upper 5 km and that this creep rate can be locally high (e.g., 8–10 mm/yr). The creep rate estimates in the upper 5 km do not appear to be sensitive to choice of locking depth. For example, holding the locking depths and dips inferred from the GPS data using Bayesian model class selection (Table S5) fixed in a Bayesian inversion for creep rates and deep slip rates results in creep rate estimates for the MF (above 5 km) and BSF (above the locking depth of 4 km) that are essentially the same as those of Figure 6 which were obtained when the locking depths for these two faults were fixed at the values based on the relocated seismicity of Waldhauser and Schaff [2008].

From 5 km depth to the locking depth, the estimated creep rate on the MF is 2.9 ± 3.5 mm/yr. This suggests that the MF is nearly locked below 5 km, in agreement with evidence for 2–3 surface-rupturing earthquakes on the MF since 770 C.E. (C. S. Prentice et al., submitted manuscript, 2014). The MF creep rate estimate for the 5 km to 9 km depth range does not trade off strongly with MF creep rate estimates above 5 km but is anticorrelated with the fault's deep slip rate estimate (Figure 11). This implies that the GPS data could be fit with somewhat more creep from 5 to 9 km and a corresponding decrease in deep slip rate. However, at 95% confidence the upper bound creep rate from 5 to 9 km (6.4 mm/yr) is still only ~50% of the 95% confidence lower bound on deep slip rate (12.1 mm/yr).

Our estimated BSF creep rate from 5 km to 13 km depth is 7.7 ± 2.4 mm/yr from which we infer that the BSF is fully creeping below 5 km (or, restated, that the locking depth is shallow). However, because our model treats the creep rate from 5 km to 13 km depth as uniform along strike, we cannot rule out the existence of localized locked zones that could give rise to moderate earthquakes. The BSF creep rate estimate for the

5–13 km depth range trades off more with the estimated creep rates in the upper 5 km than it does with the BSF deep slip rate (Figure 11).

The dip-slip creep rate estimates are generally less than half the strike-slip creep rate estimate for the same location on the fault, and most of the dip-slip creep rate estimates are not significantly different from zero at two sigma. While the inferred dip-slip creep rates for both the MF and BSF show a mixture of reverse and normal motion, the MF tends toward reverse creep and the BSF toward normal creep in our results. This is in keeping with evidence for reverse faulting in the MF zone reported by C. S. Prentice et al. (submitted manuscript, 2014) in their Willits trenching study and focal mechanisms reported by *Castillo and Ellsworth* [1993]. Extensive landsliding obscures fault-related geomorphic features along much of the BSF [Lienkaemper, 2010], limiting the extent to which correlations between such features and inferred dip-slip creep may be drawn. The prominent right step at Lake Pillsbury forms a pull-apart basin bounded on the northeast by a normal fault with evidence for Holocene movement [Lienkaemper, 2010]. However, GPS velocities are nearly fault parallel at this location, and the inferred dip-slip creep is nearly pure strike slip. The complexity of the BSF surface trace which has numerous bends and stepovers [Lienkaemper, 2010], along with the lack of a clearly defined MF or BSF plane in the relocated seismicity, might be expected to give rise to localized variations in sense of motion on the shallow creeping portion of these faults.

6.2.2. Comparison With Other Creep Observations

The creep rate estimates can be compared to surface creep measured using alinement arrays [McFarland et al., 2009]. The locations of these arrays within our study area are shown in Figure 1, and the creep rate comparisons discussed in the following paragraphs are shown in Figure 3. There are three alinement arrays on the portion of the MF included in our creep model. The southeasternmost array (named MAMR) is located at Middle Ridge and has been measured for 4 years. The average creep rate estimated from these data is 2.0 ± 3.4 mm/yr (all uncertainties for creep rates from alinement array data are stated at the 2 sigma level). Our creep rate estimate of 4.9 ± 1.2 mm/yr on the upper 5 km of the fault at this location is consistent with the surface creep rate measurement.

The Ukiah alinement array (MUKI) has recorded an average creep rate over 19 years of 4.3 ± 1.5 mm/yr. Our creep rate estimate at this location of 1.8 mm/yr ± 2.6 mm/yr is consistent with the surface creep rate measured by the alinement array at 95% confidence. The relative velocities of near-fault pairs of GPS sites that span the MF near Ukiah are consistent with a near-surface creep rate of ~ 5 mm/yr (Figure 10a, Green's function 5). Two factors may lead to the apparent underestimate of creep rate in our model at this location. First, summing the Green's functions for subfaults in the upper 5 km of the MF effectively coarsens the fault geometry, and the creep rate estimated with the Bayesian inversion is the average rate over a much larger area of the fault than that to which the alinement array is sensitive. Second, the velocities for CGPS sites have much lower uncertainties than the campaign velocities and therefore exert more control over the solution. In our model, the fit to the data at P190 (see Figure 1 for location), in particular, is sensitive to increased creep in the vicinity of Ukiah.

The average creep rate measured at the Willits alinement array (MWIL) over 20 years is 5.7 ± 0.2 mm/yr [McFarland et al., 2009]. Our results show an estimated creep rate of 9.9 mm/yr at this location with a lower 95% confidence limit of 8.3 mm/yr. This alinement array is installed on the main mapped MF trace but may not span the fault zone completely (J. Lienkaemper, personal communication, 2013). Furthermore, any MF deformation accommodated by the East Willits Fault (C. S. Prentice et al., submitted manuscript, 2014) or other active strands would not be recorded by this array. Therefore, the Willits alinement array might be expected to record a surface creep rate that is lower than the average creep rate we estimated in the upper 5 km. The GPS residual vectors in this area do not show a systematic orientation that would indicate mismodeled creep rate (Figure 9).

Three alinement arrays have been installed on the BSF in our study area [McFarland et al., 2009]. Two of these arrays, at Highway 20 and Newman Springs, span the fault at locations that correspond to an estimated creep rate of 4.6 ± 2.6 mm/yr from the Bayesian inversion. The surface creep rate recorded at the Highway 20 array (BS20) of 0.4 ± 1.8 mm/yr over 2.4 years is consistent with our estimate of 2 sigma; however, at least 5 years of observations are typically required to obtain a reliable creep rate estimate from alinement array data (J. Lienkaemper, personal communication, 2013). The Newman Springs alinement array has recorded left-lateral motion over 3.8 years but as noted by McFarland et al. [2009] is not reliable due to an unstable benchmark. Profiles of fault-parallel GPS velocities (Figure 10b, Green's function 8 and Green's

function 9) suggest that the surface creep rates may decrease as one moves northwest from Highway 20, in agreement with our estimate of no creep in the upper 5 km on the BSF for several kilometers along strike northwest of the Newman Springs alignment array.

The alignment array at Lake Pillsbury (BSLP) has the longest observing history of the BSF arrays and has recorded an average surface creep rate of 3.1 ± 0.4 mm/yr over 6.6 years [McFarland *et al.*, 2009]. This is in very good agreement with the fault-parallel velocities of GPS sites spanning the fault at this location (Figure 10b, Green's function 10). The estimated creep rate in the upper 5 km near Lake Pillsbury obtained from the Bayesian inversion is 4.3 ± 1.6 mm/yr, consistent with the alignment array measurement at 2 sigma. The creep recorded at the Lake Pillsbury alignment array exhibits episodic behavior with periods of quiescence punctuated by creep events; a longer observing history will permit a more precise estimate of the time-averaged surface creep rate.

Tong *et al.* [2013] analyzed ALOS PALSAR interferograms for the 2006–2010 time period to obtain line of sight velocities. From these they derived surface creep rate estimates as a function of along-strike location for faults throughout the San Andreas system. As with the results of the Bayesian inversion, Tong *et al.*'s estimated creep rates for the MF and BSF show substantial (although not necessarily significant) magnitude variation over relatively short distances (~10 km). However, many of the creep rates reported by Tong *et al.* [2013] have large uncertainties and/or are negative. Negative values could imply left-lateral creep but may also be caused by vertical motion which is not accounted for in their analysis. For the MF their creep rate estimates (excluding negative values) range from 0.2 to 8.5 mm/yr. At Ukiah they estimate a rate of 2.7 mm/yr, greater than the creep rate we inferred but less than that recorded at the Ukiah alignment array [McFarland *et al.*, 2009]. At Willits Tong *et al.* [2013] estimate a creep rate of 0.8 mm/yr, less than both our estimate and that of the Willits alignment array [McFarland *et al.*, 2009]. The Tong *et al.* [2013] creep rate estimate for the BSF at Lake Pillsbury is negative, but at the latitude of GPS station pair WOTI/ISLE their InSAR-derived creep rate estimate is 6.9 mm/yr, in good agreement with the relative velocity of these two GPS sites (Figure 10b, Green's function 11).

6.3. Implications for Moment Deficit and Hazard Assessment

6.3.1. Incorporation of Creep Into Hazard Assessment

As discussed in Weldon *et al.* [2013], prior to UCERF3, some California seismic hazard assessments incorporated the effect of fault creep by reducing the fault area used to calculate a creeping fault's moment deficit rate. The amount by which the area was reduced was determined from the ratio of the surface creep rate to the deep slip rate without accounting for the possibility that creep rate may decrease with depth. The implicit assumption that the ratio of creep rate to deep slip rate is constant from the surface to the locking depth means that the moment deficit rates (and thus hazard) might be underestimated if the creep rate in fact does decrease with depth.

To address this in UCERF3, a new approach was taken in which creep rate versus depth profiles were developed following the methodology of Savage and Lisowski [1993] [Weldon *et al.*, 2013]. This approach used the observed surface creep rate, the locking depth, and the deep slip rate to calculate the creep rate (and thus slip deficit rate relative to the assumed deep slip rate) as a function of depth. By integrating the slip deficit rate from the Earth's surface to the locking depth (for a given unit of distance along strike), they constructed a curve relating the observed quantity, the ratio of surface creep rate to deep slip rate, to the quantity required for hazard assessment, the moment deficit rate reduction. With the exception of the end-member cases in which the fault is fully locked or fully creeping, the moment deficit rate reduction will be less than the surface creep rate to deep slip rate ratio.

However, they found that in practice the results of Savage and Lisowski [1993] only apply for creep rates $\leq 50\%$ of the deep slip rate. For faults with higher creep rates the UCERF3 working group extrapolated the creep rate versus depth curves [Weldon *et al.*, 2013]. To do so they adopted the assumption that creep rate decreases from the Earth's surface downward, reaching a minimum in the 5–8 km depth range before beginning to increase with depth until reaching the deep slip rate at the locking depth. This conceptual model is based on two previous studies of the spatial distribution of creep, Murray *et al.* [2001] and Schmidt *et al.* [2005], both of which applied spatial smoothing to the inferred creep rate distribution to regularize the inversion.

The results of the Bayesian inversions that we conducted to estimate the creep rate distributions for the MF and BSF, while not disproving the conceptual model used by UCERF3, demonstrate that the spatial resolution of creep rate at depth is substantially more limited than would be implied by the results obtained with spatial smoothing. In particular, the apparent increase in creep rate from 10 km depth to the locking depth (14 km) on the Parkfield section of the San Andreas evident in the results of *Murray et al.* [2001], as well as other studies at Parkfield such as *Harris and Segall* [1987], is almost certainly due to the effects of smoothing the estimated creep rates above 14 km into the deep slip rate below 14 km. Indeed, regardless of inversion approach, the ability of geodetic data to resolve spatially variable slip at depth on a near-vertical fault is limited.

In other ways, the results of our Bayesian inversion are compatible with the UCERF3 approach to accounting for fault creep. For example, *Weldon et al.* [2013] assume that a fault whose surface creep rate is comparable to its deep slip rate creeps at that rate at all depths. Although the BSF surface creep rate varies along strike, this assumption appears to be consistent with our BSF creep rate and deep slip rate estimates. While we do not estimate the depth extent of creep explicitly, we do infer an average creep rate on the MF above 5 km that is less than half the deep slip rate but is higher than the average creep rate from 5 km to the locking depth. This is qualitatively consistent with the findings of *Weldon et al.* [2013] that for their assumed values of locking depth (taken to be 12 km everywhere) and deep slip rate, a fault whose surface creep rate is ~50% of its deep slip rate has a depth extent of creep that is about half the fault's locking depth.

Earthquakes on faults which exhibit a relatively high interseismic creep rate, even if creep does not extend to all depths, have been observed to be followed by afterslip comparable in magnitude to the coseismic slip [e.g., *Langbein et al.*, 2006; *Murray-Morealeda and Simpson*, 2009]. This behavior has also been demonstrated in earthquake simulations that include the effects of fault creep [e.g., *Aagaard et al.*, 2012]. Furthermore, recent numerical simulations by *Noda and Lapusta* [2013] suggest that velocity strengthening portions of faults that creep interseismically may still participate in coseismic ruptures under certain conditions. The possibility of rapid and large postseismic surface displacements or triggered coseismic offsets should be considered in plans for earthquake response, loss mitigation, and deformation monitoring along the MF and BSF.

6.3.2. Moment Deficit on the Maacama Fault

Our results suggest that a slip deficit is accumulating on the seismogenic zone of the MF. The existence of a slip deficit may be a necessary but not a sufficient condition for earthquake rupture [*Murray and Segall*, 2002], and individual earthquakes may not recover the full slip deficit (see, for example, *Konca et al.* [2008]). However, moment deficit rates calculated from geologic and, more recently, geodetic data are a major ingredient in seismic hazard assessment [*Field et al.*, 2014]. Therefore, here we consider our results in comparison to those of C. S. Prentice et al. (submitted manuscript, 2014), which are based on geologic data, as they pertain to slip deficit and the possible size of future earthquakes on the MF.

The slip deficit since the most recent earthquake in the MF paleoseismic record may be calculated at points along the fault using our deep slip rate and creep rate estimates and the appropriate elapsed time. C. S. Prentice et al. (submitted manuscript, 2014) present evidence for a surface-rupturing earthquake at Willits between 770 and 890 C.E. based on an offset stream channel exposed in trenches. A folding event also observed in their trenching study may represent a more recent earthquake (890–1150 C.E.). *Sickler et al.* [2005] conducted a paleoseismic study from which they concluded that the most recent surface-rupturing event on the MF at Ukiah occurred between 1410 and 1660 C. S. Prentice et al. (submitted manuscript, 2014) do not rule out the possibility of that event having ruptured as far northwest as Willits.

We calculate slip deficits at Willits based on our results for the three time periods described above and compare them to the analogous estimates obtained from the results of C. S. Prentice et al. (submitted manuscript, 2014) in Table 1. As can be seen, the estimated slip deficit based on our results is substantially higher than that predicted from the geologic slip rate estimate. This arises due to two factors. First, C. S. Prentice et al.'s slip rate estimate (6.4–8.6 mm/yr) is lower than our deep slip rate estimate of 13.1 mm/yr. Second, they assume that the surface creep rate measured at the Willits alignment array (5.6–5.8 mm/yr [*McFarland et al.*, 2009]) extends to all depths, thus reducing the slip deficit. The results of our Bayesian inversion suggest that the MF creep rate actually decreases with depth, leading to a greater slip deficit.

Table 1. Comparison of Geodetic and Geologic Estimates of Slip Deficit at Willits

Time Span Start Time (C.E.) ^a	Shallow Geodetic Slip Deficit (m) ^b	Deeper Geodetic Slip Deficit (m) ^c	Geologic Slip Deficit (m) ^{d,e}	M_w [W&C, 1994] ^f	M_w (Spatially Variable Slip Deficit) ^g	Start Time Description
1410–1660	1.1–1.9	3.6–6.2	0.2–1.8	7.1–7.3	7.0–7.1	Ukiah surface-rupturing event [Sickler et al., 2005]
890–1150	2.8–3.6	8.8–11.5	0.5–3.4	7.4–7.5	7.2–7.3	Willits folding event (C. S. Prentice et al., submitted manuscript, 2014)
770–890	3.6–4.0	11.5–12.7	0.7–3.7	7.5–7.6	7.3	Willits surface-rupturing event (C. S. Prentice et al., submitted manuscript, 2014)

^aAll slip deficit calculations use elapsed time calculated as the number of years between the dates in column 1 and 2013 C.E.

^bGeodetic slip deficit above 5 km depth calculated from the difference between this study's MF deep slip rate estimate of 13.1 mm/yr and the MF creep rate estimate at Willits (9.9 mm/yr).

^cGeodetic slip deficit above 5 km to 9 km depth calculated from the difference between this study's MF deep slip rate estimate of 13.1 mm/yr and the MF creep rate estimate at Willits (2.9 mm/yr).

^dGeologic slip deficit calculated from geologic slip rate estimate of C. S. Prentice et al. (submitted manuscript, 2014) (6.4–8.6 mm/yr) and the alignment array creep rate estimates of McFarland et al. [2009] (5.6–5.8 mm/yr).

^eMinimum values are calculated from lower slip deficit rates and shorter time spans; maximum values are calculated from higher slip deficit rates and longer time spans.

^f M_w estimates calculated using geodetic slip deficit in column 2 and scaling relationship of Wells and Coppersmith [1994].

^g M_w estimates calculated from spatially variable slip deficit derived from geodetic deep slip rate and creep rate estimate for MF between Ukiah and Willits (this study).

Wells and Coppersmith [1994, Table 2B] present scaling relationships relating an earthquake's moment magnitude (M_w) to both its average and maximum displacement at the Earth's surface. Treating a surface offset as an average, rather than the maximum, displacement results in a higher M_w estimate from these relationships. Column 5 of Table 1 gives the M_w calculated using Wells and Coppersmith's [1994] scaling relationships assuming the shallow slip deficits we estimated (Table 1, column 2) are the average surface displacements expected in a future earthquake. Treating calculated slip deficit as earthquake surface offset implicitly assumes complete recovery of slip deficit in that earthquake.

For comparison to the M_w calculated using the scaling relationship of Wells and Coppersmith [1994], we use the spatially variable creep rates for the portion of the MF between Ukiah and Willits and the MF deep slip rate that we inferred from the Bayesian inversion to calculate the moment deficit for the three time periods. From this we obtain the expected M_w for earthquakes on this section of the MF assuming a shear modulus of 30 GPa and complete recovery of all slip deficit to a depth of 9 km (Table 1, column 6).

With the exception of the longest time period given in Table 1, the anticipated M_w calculated using results from GPS inversions and either scaling relationships or inferred spatially variable slip deficit is compatible with the magnitude range for earthquakes involving the MF used in the UCERF3 analysis [Field et al., 2014]. If the most recent moderate to large earthquake on the MF was the Willits surface-rupturing event identified by C. S. Prentice et al. (submitted manuscript, 2014), the moment deficit predicted from our results would be equivalent to a M_w 7.5–7.6 earthquake. Given the diffuse nature of the MF relocated seismicity compared to that of more mature faults like the San Andreas and Hayward [Waldhauser et al., 2004; Waldhauser and Schaff, 2008], as well as the Maacama Fault's relatively recent inception [McLaughlin et al., 2012], it seems unlikely that the MF has repeatedly produced earthquakes of this magnitude. Rather, it may more typically participate in smaller events, like the Ukiah surface-rupturing earthquake, some of which may be missing from the paleoseismic record of a single fault strand.

7. Conclusions

In order to quantify the impact of fault creep on moment deficit for faults of the northern San Andreas system, we established a dense network of SGPS sites to supplement more broadly distributed CGPS sites in the region, estimated interseismic velocities from the GPS data, and used these in a Bayesian inversion to estimate deep slip rates on the SAF, MF, and BSF and spatially variable creep rates on the latter two. The Bayesian analysis allows for arbitrary prior information and fully characterizes the solution space for a given model class by providing the full posterior PDF of model parameters.

Our results suggest that creep rates on the upper 5 km of both the BSF and MF vary along strike and locally can approach the faults' deep slip rates. Creep on the MF occurs primarily above 5 km, implying that the MF is accumulating moment deficit. The inferred BSF creep rate from 5 km to 13 km depth is comparable to its deep slip rate. One interpretation is that the BSF locking depth is substantially shallower than the depth extent of microseismicity. However, we cannot rule out the existence of locked patches on the BSF in the 5–13 km depth range that are of a spatial extent that is not resolvable with the available data.

To assess the extent to which fault creep might reduce seismic hazard by limiting moment deficit, the creep rate estimates must be evaluated in the context of the faults' deep slip rates. Our SAF deep slip rate estimate agrees with geologic rate estimates, and there is no geologic estimate to which to compare our BSF deep slip rate. For the MF, however, our deep slip rate estimate is roughly twice the geologic slip rate estimate at Willits. This, combined with the inference from GPS data that the MF surface creep rate does not extend to all depths, results in an estimated moment deficit that is correspondingly larger than what has been inferred geologically. Possible explanations for the higher deep slip rate inferred from geodesy include the existence of multiple active fault strands that are spanned by the GPS but not the geologic observations or distributed off-fault strain that is mapped to fault slip in the geodetic inversions.

Several lines of follow-on work could help refine the results presented here. The trade-off between locking depth and deep slip rate is a major source of epistemic uncertainty in geodetically constrained dislocation models. While this uncertainty is not reflected in the posterior PDFs obtained in our analysis, development of approaches for the general inclusion of epistemic uncertainty in Bayesian inversion has the potential to address this need in the future. Extension of models to incorporate more complex features of the regional geology, for example, the structural fabric and rheological signature left by the passage of the Mendocino triple junction, might provide additional useful insights into partitioning of deformation in the northern San Andreas system. Additional geodetic data collection between the MF and SAF might help constrain the SAF deep slip rate and locking depth, although the lack of off-shore data will remain a serious limitation. Continued efforts to obtain geologic slip rate estimates for the BSF could result in valuable independent information to which the geodetic rate estimates could be compared.

8. Data and Resources

CGPS data used in this study were recorded by instruments of the Plate Boundary Observatory (PBO, <http://pbo.unavco.org/>) and the Bay Area Regional Deformation (BARD) network (<http://seismo.berkeley.edu/bard/>). The PBO is operated by UNAVCO for EarthScope (www.earthscope.org) and supported by the National Science Foundation (No. EAR-0350028 and EAR-0732947). The majority of campaign GPS data used here was collected by the USGS. Additional GPS data collected by Stanford University [Frey Mueller *et al.*, 1999] and UNAVCO were obtained from the UNAVCO Facility archive which is maintained with support from the National Science Foundation (NSF) and National Aeronautics and Space Administration (NASA) under NSF Cooperative Agreement No. EAR-0735156. Processed GPS time series used in this study are available at <http://earthquake.usgs.gov/monitoring/gps/SFBayArea> and http://earthquake.usgs.gov/monitoring/gps/NCalifornia_SGPS. Alinement array data were obtained from <http://pubs.usgs.gov/of/2009/1119/>. Mapped fault traces were obtained from <http://pubs.usgs.gov/ds/541/> and Dawson [2013]. Maps were generated using Generic Mapping Tools [Wessel *et al.*, 2013].

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