Using Surface Creep Rate to Infer Fraction Locked for Sections of the San Andreas Fault System in Northern California from Alignment Array and GPS Data


Abstract  Surface creep rate, observed along five branches of the dextral San Andreas fault system in northern California, varies considerably from one section to the next, indicating that so too may the depth at which the faults are locked. We model locking on 29 fault sections using each section’s mean long-term creep rate and the consensus values of fault width and geologic slip rate. Surface creep rate observations from 111 short-range alignment and trilateration arrays and 48 near-fault, Global Positioning System station pairs are used to estimate depth of creep, assuming an elastic half-space model and adjusting depth of creep iteratively by trial and error to match the creep observations along fault sections. Fault sections are delineated either by geometric discontinuities between them or by distinctly different creeping behaviors. We remove transient rate changes associated with five large (M ≥5.5) regional earthquakes. Estimates of fraction locked, the ratio of moment accumulation rate to loading rate, on each section of the fault system provide a uniform means to inform source parameters relevant to seismic-hazard assessment. From its mean creep rates, we infer the main branch (the San Andreas fault) ranges from only 20%–10% locked on its central creeping section to 99%–100% on the north coast. From mean accumulation rates, we infer that four urban faults appear to have accumulated enough seismic moment to produce major earthquakes: the northern Calaveras (M 6.8), Hayward (M 6.8), Rodgers Creek (M 7.1), and Green Valley (M 7.1). The latter three faults are nearing or past their mean recurrence interval.

Online Material: High-resolution fault system map, and tables of creep observations and geographic coordinates of fault model.

Introduction

The dextral northern San Andreas fault system (NSAFS; Fig. 1) extends from Parkfield, California, (Fig. 2) northward to the Mendocino triple junction and consists of five major branches that combine for a total linear extent of nearly 2000 km. We estimate that approximately 60% of the fault system’s total length exhibits significant seismic moment release through fault creep. However, because many creeping fault sections have substantial locked areas at depth (e.g., Lienkaemper et al., 2012), only about 28% of the system’s total seismic moment budget is released aseismically by fault creep. Thus, developing a better understanding of how the depth extent of creep varies on the creeping sections is fundamental to regional seismic-hazard assessment. This report describes a uniform approach that uses only long-term surface creep data and consensus fault parameters to estimate the locked fraction of each of 29 fault sections. The locked fraction allows us to infer the rate of seismic moment accumulation per section. We evaluate the potential size of future earthquakes from the estimated accumulated seismic moment over the time elapsed since the most recent earthquake. We then compare the amount of elapsed time with the mean earthquake recurrence time for each of the major urban creeping faults in the San Francisco Bay region, where we also have historical and paleoseismic information.

Since the previous summary of surface creep rates in this region, primarily from alignment array data (Galehouse and Lienkaemper, 2003), we have greatly expanded the extent and density of our creep monitoring, especially north of San Francisco Bay (McFarland et al., 2014). Recently records of up to a decade have become available for continuous Global Positioning System (cGPS) stations near many sections of the NSAFS (see U.S. Geological Survey [USGS], Data and Resources). These longer GPS records allow us to infer surface creep rates to compare with and to augment the
alignment array estimates of average surface creep rate along each fault section used in this modeling. When long-term records are available from survey (campaign)-mode GPS (sGPS), these records can also be used, especially where records are long or can include a cGPS station in the pair to decrease the error. The intent is to average the best available
data of all types (excluding Interferometric Synthetic Aperture Radar [InSAR] for the reasons explained below) to optimally characterize the long-term, interseismic surface creep rate for each of the 29 fault sections in the model (Figs. 1, 2; Table 1).

To simplify complex fault nomenclature, we describe the NSAFS as south-to-north numbered sections of four branching zones related to the primary fault branch, the San Andreas (B in Fig. 1). The main focus is on the two next largest branches (C and D), which tend to exhibit significant creep on most sections and have the next highest long-term slip rates (Table 1) and the greatest overall branch lengths within the region of investigation (Fig. 1; Table 1). Branch C begins as the southern and central Calaveras fault (section C1) by stepping and branching from the central creeping section of the San Andreas fault (section B2); it continues by stepping left and branching (at depth) to the Hayward fault (sections C3–C6; Simpson et al., 2004). Section C2, the southeast extension of the Hayward fault (Fig. 2), is not used in the model for simplicity, as explained in the Analysis section. Near where it branches left to the Hayward fault, section C1 also branches right at an extensional bend that begins branch D as the northern Calaveras fault (D1–D2). In Figures 1 and 2, we plot the branches of the NSAFS using a Hotine oblique projection centered near the north tip of and subparallel to the Hayward fault (35° N). This projection was chosen to approximate the grid used in our fault model, which is based on the Sierra–Pacific pole (−94.6°E, 46.7°N) of Argus and Gordon (2001); this pole is derived from GPS velocities, with its origin (i.e., central meridian, −122.366803°E, base latitude 38.002942° N) coinciding with the 0 km origin of the Hayward fault grid of Lienkaemper (1992, 2006). Using a grid aligned with the plate boundary gives a synoptic view of the variation in creep rate and other properties along the fault branches (e.g., Fig. 3). We refer to all locations along each of the five fault branches, including arrays and GPS stations, as kilometer distances along this plate boundary grid referenced to this common 0 km origin for the x direction.

In the following sections, we describe our survey methods, the surface creep observations, and the very simple elastic model used to estimate variations in the depth of creep and percentage of the seismic moment accumulation (i.e., fraction locked) for each of the 29 fault sections used in the model. We first summarize results for each of the five fault branches, and then for the fault system as a whole. Finally, we discuss possible implications of the results for seismic hazard on urban creeping fault sections and directions toward future improvements in both creep monitoring and modeling of the data.
Table 1
Mean Surface Creep Rate on Sections of the Northern San Andreas Fault System

<table>
<thead>
<tr>
<th>Fault ID*</th>
<th>Fault Section</th>
<th>Observed Surface Creep Rate (mm/yr)</th>
<th>Fault Section Valuesa</th>
<th>Depth of Creep (km)</th>
<th>Fraction Locked (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Mean (mm/yr)</td>
<td>±1 SEM</td>
<td>N used</td>
<td>N GPS</td>
</tr>
<tr>
<td>A1</td>
<td>San Gregorio</td>
<td>0.1</td>
<td>0.4</td>
<td>2</td>
<td>1</td>
</tr>
<tr>
<td>B1</td>
<td>San Andreas, south transition</td>
<td>10.5</td>
<td>1.8</td>
<td>8</td>
<td>2</td>
</tr>
<tr>
<td>B2</td>
<td>San Andreas, central creeping</td>
<td>25.1</td>
<td>1.0</td>
<td>13</td>
<td>3</td>
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<tr>
<td>B3</td>
<td>San Andreas, north transition</td>
<td>13.3</td>
<td>0.6</td>
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<td>1</td>
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<td>B4</td>
<td>San Andreas, SJB-GG</td>
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<td>0.4</td>
<td>6</td>
<td>2</td>
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<td>B5</td>
<td>San Andreas, GG-MTJ</td>
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<td>0.2</td>
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<td>1</td>
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<tr>
<td>C1</td>
<td>Calaveras</td>
<td>10.2</td>
<td>0.7</td>
<td>7</td>
<td>3</td>
</tr>
<tr>
<td>C2</td>
<td>Hayward, southeast extension</td>
<td>–</td>
<td>–</td>
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<td>1</td>
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<td>C3</td>
<td>Hayward, south</td>
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<tr>
<td>C4</td>
<td>Hayward, central</td>
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<td>17</td>
<td>2</td>
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<tr>
<td>C5</td>
<td>Hayward, north</td>
<td>5.4</td>
<td>0.3</td>
<td>4</td>
<td>1</td>
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<tr>
<td>C6</td>
<td>Hayward, San Pablo Bay</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>1</td>
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<tr>
<td>C7</td>
<td>Rodgers Creek</td>
<td>1.5</td>
<td>0.3</td>
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<td>1</td>
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<tr>
<td>C8</td>
<td>Maacama, south</td>
<td>0.7</td>
<td>0.7</td>
<td>1</td>
<td>–</td>
</tr>
<tr>
<td>C9</td>
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<td>8</td>
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<td>C10</td>
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<td>0.1</td>
<td>1</td>
<td>3</td>
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<td>C11</td>
<td>Briceland–Bear River</td>
<td>4.0</td>
<td>0.4</td>
<td>1</td>
<td>1</td>
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<tr>
<td>D1</td>
<td>North Calaveras, south</td>
<td>4.4</td>
<td>0.1</td>
<td>1–</td>
<td>1</td>
</tr>
<tr>
<td>D2</td>
<td>North Calaveras, north</td>
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<td>0.3</td>
<td>5</td>
<td>1</td>
</tr>
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<td>11</td>
<td>4</td>
</tr>
<tr>
<td>D4</td>
<td>Berryessa, east trace</td>
<td>0</td>
<td>1.3</td>
<td>1–</td>
<td>1</td>
</tr>
<tr>
<td>D5</td>
<td>Hunting Creek</td>
<td>1.8</td>
<td>0.2</td>
<td>2</td>
<td>1</td>
</tr>
<tr>
<td>D6</td>
<td>Bartlett Springs, Highway 20</td>
<td>2.5</td>
<td>1.0</td>
<td>2</td>
<td>1</td>
</tr>
<tr>
<td>D7</td>
<td>Bartlett Springs, Rough</td>
<td>–</td>
<td>0.2</td>
<td>3</td>
<td>3</td>
</tr>
<tr>
<td>D8</td>
<td>Bartlett Springs, Lake Pillsbury</td>
<td>2.4</td>
<td>0.7</td>
<td>2</td>
<td>1</td>
</tr>
<tr>
<td>D9</td>
<td>Bartlett Springs, Round Valley</td>
<td>3.3</td>
<td>1.7</td>
<td>5</td>
<td>3</td>
</tr>
<tr>
<td>D10</td>
<td>Lake Mountain–Eaton’s Rough</td>
<td>1.9</td>
<td>0.4</td>
<td>2</td>
<td>–</td>
</tr>
</tbody>
</table>

Notes: Mean fits 2 best rates
Consider high pre-1989 rates
Only rate near south end
GPS (1–3 yr) rates not used
AA and GPS differ considerably
SEM on 3-week GPS rates
Minimum, GPS: WOTI–ISLE
Both GPS and AA weak
(continued)
Creep Observations

Alignment Array Data

The primary source of surface creep data is the alignment array measurements made by the San Francisco State University Creep Project covering the greater San Francisco Bay region that was begun in 1979 (Galehouse and Lienkaemper, 2003; McFarland et al., 2014) and currently includes 79 active arrays (Figs. 2, 3; Table S1 available in the electronic supplement to this paper). Galehouse and Lienkaemper (2003) includes a comprehensive description of the history of discovery of creep in this region, early monitoring efforts, and a more detailed description of surveying procedures. Here, this method is only briefly summarized by referring to an inset in Figure 1 that shows a typical alignment array. A theodolite instrument is set up with a mark on one side of the fault (IS, instrument station) in sight of a target set up across the fault over another mark (ES, end station). Angles are read to the ES mark with respect to a second target set over an orientation station (OS) mark. The angle (ES–IS–OS) is read eight times, including a 180° rotation of targets and instrument, in both the forward and reversed telescope positions, and error checks on readings to prevent mistakes or excessive instrument drift. The mean angle is compared with subsequent measurements by trigonometrically converting angle differences to slip. Survey error generally tends to increase with line length, and angle estimation becomes less accurate when targets are closer than 60 m. In practice, ~100 m length arrays tend to result in the lowest error. The actual range of array lengths varies from 50 to 450 m. Shorter arrays can have more risk of missing some of the deformation zone. An ideal array crosses nearly perpendicular to a single narrow fault trace, as inferred from geomorphic or other evidence of recent movement. These data are updated annually (McFarland et al., 2014), and those used for this analysis are tabulated in Table S1.

We also use additional alignment array data, mostly older arrays referenced in Galehouse and Lienkaemper (2003) (and in Table S1) with their locations shown in Figure 2. Along the creeping sections (B1–B3) of the San Andreas fault, Burford and Harsh (1980, hereafter referred as BH80) published creep rates for many arrays (Figs. 2a, 3) but without error estimates. We approximate their error to be ∼1 mm/yr (1σ), given the relatively brief duration of these surveys (3–10 yr). Titus et al. (2006) reoccupied some of the BH80 arrays, finding somewhat lower rates for the longer period. Following the 2004 Parkfield earthquake, Lienkaemper et al. (2006) re-examined seven arrays along the Parkfield section (B1) and computed their long-term interseismic creep rates. A few older, small-aperture trilateration arrays referenced in Galehouse and Lienkaemper (2003) and Table S1 still represent the best available creep rates in some locations where newer data are lacking (Figs. 2 and 3), but most have been superseded by more accurate longer-term data. Where many measurements are available, such as for sections B1–B3, some

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### Table 1 (Continued)

<table>
<thead>
<tr>
<th>Fault Section</th>
<th>Observed Surface Creep Rate (mm/yr)</th>
<th>Depth of Creep (km)</th>
<th>Fraction Locked (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Mean (±1 SDM) used</td>
<td>Min (&lt;−2 SDM) Mean +/−2 SDM</td>
<td>Max (+2 SDM) Mean +/−2 SDM</td>
</tr>
<tr>
<td>E1 Greenville, south</td>
<td>0.5 0.7 1</td>
<td>1 1 1</td>
<td>20 15 15</td>
</tr>
<tr>
<td>E2 Greenville, central</td>
<td>0.6 0.4 1</td>
<td>1 1 1</td>
<td>20 15 15</td>
</tr>
<tr>
<td>E3 Greenville, north</td>
<td>2.1 0.4 3</td>
<td>1 1 1</td>
<td>20 15 15</td>
</tr>
</tbody>
</table>

*Fault section identifier shown in Figures 1 and S1; SJB, San Juan Bautista; GG, Golden Gate; MTJ, Mendocino triple junction (B5 is approximately located at the junction with the Mendocino fault). |

†SEM, standard error of mean of N used; AA, alignment array; T, trilateration; ‡fault section identifier shown in Figures 1 and S1; SJB, San Juan Bautista; GG, Golden Gate; MTJ, Mendocino triple junction (B5 is approximately located at the junction with the Mendocino fault). |

‡LW, width and deep slip (loading) rate values after Working Groups on California Earthquake Probabilities (2003, 2008) consensus values.
estimates deemed less reliable were excluded from the average creep rate determination for a section, indicated by the number used (Table 1) and rates marked in parentheses (Table S1).

The use of InSAR to infer creep rates of the San Andreas fault system by Tong et al. (2013) represents a promising new method that can add considerable fine detail in local fault complexity and variation in creep rate, especially in arid areas. However, north of Parkfield, both rainfall and vegetation cover increase considerably, and the creep rates determined in this InSAR study appear to be generally far noisier than most of the other data we selected. Because they have been closely integrated with cGPS data that we use as well, they do not provide fully independent evidence.

The average creep rates shown in Table 1 are intended to reflect stable long-term interseismic rates, excluding any transient rate changes associated with large earthquakes. However, some sections have had sizable changes in rate imposed by large earthquakes in the region that must be completely removed from the long-term average rate. Fault sections (C1 and C3) experienced significant changes in creep rate caused by large local earthquakes. Following the 1989 M 6.9 Loma Prieta earthquake, creep along the southern

**Figure 3.** Plots for the three largest fault branches, showing creep rate observations and modeled creep rate versus distance using the grid described in Figure 1. Fault section names are given in Table 1. (©) Data are given in Tables S1 and S2. The color version of this figure is available only in the electronic edition.
1989 (10 yrs, because most sites now exhibit rates similar to pre-mature creep rate (Table S1). The color version of this figure is available only in the electronic edition.

Ruby Canyon net (2.8 km southeast of CVCR; M. Lisowski and N. King, unpublished data, 1982). The color version of this figure is available only in the electronic edition.

The 1984 Hayward fault (C3) stopped or slowed dramatically for more than 6 yrs, attributable to an imposed shear stress drop (Lienkaemper et al., 1997, 2001, 2012). Consequently, we estimate the average creep rate (Table 1) over about the past 10 yrs, because most sites now exhibit rates similar to pre-1989 (Lienkaemper et al., 1991, 1997). However, not all sites in the fastest creeping part of this section (63–66 km), formerly creeping at ~9 mm/yr, have yet resumed their earlier fast rates. To reflect that the current mean creep rate is still much lower than the pre-1989 rate for 63–66 km, we adjust the C3 section mean rate used in solving for depth of creep by 0.3 mm/yr. That is, we solve for mean depth of creep on C3 by fitting to an adjusted mean creep rate of 7.2 mm/yr, rather than 6.9 mm/yr, the currently observed mean rate for C3 (Table 1).

Figure 3 shows that using this adjustment does fit most current observations reasonably well, and we consider it an overall better fit than not having made an adjustment.

Figure 4. Corrections to long-term creep rate of two sites in fault section C1 (CVCR and CVWR; Fig. 2a, Table S1) for the effects of large local earthquakes. Earthquake magnitudes and dates are shown by vertical lines. Steps in the accumulated creep are indicated by arrows calculated using multiple linear regression as described in the text. Ruby Canyon net (2.8 km southeast of CVCR; M. Lisowski and N. King, unpublished data, 1982). The color version of this figure is available only in the electronic edition.

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Two sites on the Calaveras section (C1) monitored the longest, CVCR at ~128 km and CVWR at ~152 km (Figs. 2, 4), have been significantly affected by multiple large local earthquakes, and thus the background long-term creep rate must exclude these effects. Surveying began on array CVCR in 1968 before the 1979 M 5.9 Coyote Lake earthquake, which ruptured at depth below it (Oppenheimer et al., 1990), and rapid afterslip of ~8 cm occurred in that area over about six months. Figure 4 includes a more complete afterslip record measured on the Ruby Canyon net 2.8 km to the south. The 1984 M 6.2 Morgan Hill earthquake rupture ended just north of the 1979 rupture but ultimately released somewhat greater slip (~14 cm) at CVCR as a transient episode lasting nearly 20 yrs. The more distant M 6.9 Loma Prieta earthquake may have contributed slightly to this transient, given that Simpson and Reasenberg (1994) calculated a small dextral shear stress increase for the location of CVCR after that earthquake. However, we lack measurements from 1989 to clearly distinguish this effect from the ongoing post-1984 transient. Since mid-2004, the creep rate on CVCR is essentially identical to the ~10 mm/yr pre-1979 rate; 10.3 ± 0.3 mm/yr is obtained using multiple linear regression to jointly estimate the background rate and the earthquake-correlated steps. The other array, CVWR, first surveyed shortly after the 1979 earthquake, may have recorded a minor post-1979 transient. However, for purposes of computing the long-term background creep rate, 10.1 ± 0.1 mm/yr, only the 2 cm sinistral response from the 1989 earthquake was eliminated using multiple linear regression.

**GPS Data**

Either sGPS or cGPS velocities from stations located near creeping faults can provide an important additional constraint on surface creep rate, especially in remote regions with steep terrain, where it can be extremely difficult or impossible to install stable alignment arrays. In rugged terrain, braced GPS monuments can be placed on stable hilltops at substantially greater distances from the fault with less loss of accuracy than with an alignment array; however, the
uncertainty in modeling surface creep rates increases as distance increases. cGPS data are preferred where available, because uncertainty can generally be reduced much more quickly than for sGPS, which often requires nearly a decade of observation. However, having greater sGPS station redundancy near the fault on one section can also be useful for resolving its mean surface creep rate. USGS cGPS and sGPS cleaned, North America fixed velocity files were accessed on 6 May 2013 (USGS, see Data and Resources). These velocities and the surface creep rates derived from them are available in Table S2. The general approach we used for inferring creep rates from GPS observations on the Greenville fault was described in Lienkaemper, Barry, et al. (2013). That approach was initially based on equations of Savage and Lisowski (1993), although in this paper we have modified the approach to use slip estimates supplied by our simple elastic dislocation model (see Fig. S2 and its description in the electronic supplement for two examples).

Analysis

Our goal is to use a simple model to estimate the average depth of creep for each fault segment (Fig. 5) using the best long-term estimates of surface creep rates from alignment arrays and pairs of GPS stations. An ultimate simple model would be a vertical 2D dislocation model of the segment at depth straddled at the Earth's surface by the alignment arrays and GPS pairs with consensus fault parameters used for driving rate and depth of locking. However, as Savage and Lisowski (1993) demonstrated for the Hayward fault, deep slip on the adjacent subparallel Calaveras and San Andreas faults contributes to the forces driving creep on the Hayward fault, so we wanted to include the driving contributions of adjacent faults at least approximately. Given the many uncertainties, our goal was not to make a comprehensive model of northern California faults that could fit all the absolute GPS velocities, as desirable as that might be for other purposes.

We estimate the average depth of creep for each fault section (Fig. 5) using a model fault geometry consisting of an upper creeping (shear-stress free) patch over a locked zone. Both are loaded from below by a deep vertical dislocation slipping at a long-term slip rate or driving rate. Driving (loading) rates and fault widths (i.e., depths to bottom of locked zone) are after the Working Group on California Earthquake Probabilities (2003, 2008) with minor modifications (Table 1). Driving dislocations begin at the base of the locked zones and extend to a depth of 1000 km. To prevent abrupt changes in loading at the edges of the model, we provide long lateral extensions with deep driving dislocations for those fault branches that continue well beyond the modeled area (for fully locked patches along branch A southward and branch B northward and southward; and also for the creeping patches along branches C and D northward; all shown as dotted lines in Fig. 1). The method is similar in concept to the one used along the Hayward fault, where creep rate data are far more abundant along strike than for other faults (Savage and Lisowski, 1993; Simpson et al., 2001; Lienkaemper et al., 2012). Fault section ends are placed either at major geometric discontinuities or where creeping behavior differs significantly from adjacent sections. All sections are simply modeled as vertical planes, which ignore an additional uncertainty factor in some remote sections of branches C and D (Murray et al., 2014). We use the equations of Okada (1992) to predict surface creep rates from a 3D forward model based on estimated long-term creep rates and depths for all 29 fault sections. The electronic supplement contains a text file of all geographic coordinates for the modeled fault sections and their extensions.

We begin by making initial estimates of creep depth for the creeping patch for all fault sections, then run a boundary value dislocation solver to estimate a patch creep rate(s) on each section. Model sections simplify active surface trace geometry but are intended to capture significant changes in trend. Some sections have been divided into creeping subpatches to estimate subpatch creep rates (Fig. 5). Estimation of best-fit depths for each section was done in stepwise fashion by trial and error. To test each trial depth for its goodness of fit to observed surface creep rate, we calculated synthetic surface creep rates at the positions of all creep observation sites. Average model depth of creep was adjusted iteratively by hand for the 29 fault sections until all of the modeled mean section surface creep rates fit to within ±0.1 mm/yr of the observed mean surface creep rates.

For GPS station pairs on opposite sides of a fault trace, we estimate a surface creep rate at the position on the fault trace nearest the western station of each pair. In much of the modeled region, the easternmost block tends to be the more rigid one, with its thicker crust, coupled to the Sierra Nevada–Great Valley or Sierran microplate (Argus and Gordon, 2001).
Such GPS-derived creep rates generally agree well with nearby alignment array data (Fig. 3; Tables 1 and 2).

After obtaining the values of mean creep depth that satisfy average observed surface creep rates on all sections, we then determine the 95-percentile range of uncertainty in creep depth for each fault section by varying its creep depth until modeled surface creep rates fit the 95-percentile uncertainty range in observed surface creep rate for the section (Figs. 3 and 6). The effect of uncertainty in the driving depth and driving rates for each section is not included, partly because it presents a much more difficult computational problem. But, more importantly, by not including these uncertainties, it allows us to clearly evaluate how much of the uncertainty in the depth of creep, and of the proportion of seismic to aseismic release, is directly attributable to the calculated uncertainties or other deficiencies in current surface creep observations.

Although this manual method of fitting lacks a systematic test of sensitivity, the depth of creep estimates for adjacent creeping faults appears to be mutually insensitive, primarily because driving slip rate and depth are held fixed for all sections and most fault sections are tens of kilometers long. For example, fault section C2 (Fig. 2), the southeastern extension of the Hayward fault, was included in early runs of the model. For simplicity, we completely eliminated the C2 section from the model, along with its supposed 3 mm/yr driving rate. The earlier model had mistakenly included 3 mm/yr excess driving slip on the part of section C1 adjacent to C3. However, eliminating section C2 and its excess driving slip, caused negligible impact (≤0.1 km) on the depth of creep calculated on any of the three nearest fault sections (C1, C3, and D1), partly because C1 is such a long section. This example provides a relevant test of sensitivity, because these three sections are only a few kilometers apart, making it a logical place to test for possible trade-offs between the creep depth and driving rate variation. The electronic supplement includes a more general discussion of the sensitivity of locking estimates to uncertainty in the driving rate and depth.

Using the estimated creep depths and creep rates on the subpatches of all fault sections, we estimate how aseismic release rates and amount of locking may vary along the five branches of the NSAFS. For all creeping subpatches along each fault section, we sum the aseismic moment release rate, the pinned seismic moment in the creeping patches (i.e., loading rate less patch creep rate), and calculate moment stored in the lower locked patch. The intent is to estimate the fraction locked based on the amount of seismic moment available for release in large (M ≥ 5.5) earthquakes. The fraction locked, as used here, is intended to have the same meaning as the seismic scaling factor (R) of Bakun (2003). However, later we consider how our definition differs in actual usage, especially for those fault sections that are dominantly creeping, particularly section C1, the central and southern Calaveras fault.

Thus, fraction locked for these purposes represents the sum of (1) the seismic moment accumulation rate on the lower locked patch and (2) for simplicity, half of all the moment accumulation rate held by pinning within the creeping subpatches. This sum is divided by the seismic moment loading rate applied to the fault section (Table 1; Fig. 7). This simply assumes that only half of the strain accumulated by pinning in the creeping zone will release dynamically (i.e., true coseismic release) and that the rest will release as afterslip, which is counted as part of the aseismic release rate for this purpose. However, this overly simple assumption must have the caveat that there are only a few well-documented historical field investigations of afterslip to inform the current empirical understanding of afterslip. Existing global afterslip

<table>
<thead>
<tr>
<th>Fault Branch</th>
<th>Moment Accumulation Rate, Locked Areas (A) (N·m yr⁻¹)</th>
<th>Moment Releasing Rate, Creeping Areas (B) (N·m yr⁻¹)</th>
<th>Moment Loading Rate, A + B (N·m yr⁻¹)</th>
<th>Fraction Locked (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A, San Gregorio Mean</td>
<td>2.95 × 10¹⁷</td>
<td>0.00</td>
<td>2.95 × 10¹⁷</td>
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<tr>
<td>+2σ</td>
<td>2.70 × 10¹⁷</td>
<td>2.51 × 10¹⁶</td>
<td>2.95 × 10¹⁷</td>
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</tr>
<tr>
<td>B, San Andreas Mean</td>
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<td>1.18 × 10¹⁸</td>
<td>5.48 × 10¹⁸</td>
<td>79</td>
</tr>
<tr>
<td>−2σ</td>
<td>4.47 × 10¹⁸</td>
<td>1.01 × 10¹⁸</td>
<td>5.48 × 10¹⁸</td>
<td>82</td>
</tr>
<tr>
<td>+2σ</td>
<td>4.10 × 10¹⁸</td>
<td>1.38 × 10¹⁸</td>
<td>5.48 × 10¹⁸</td>
<td>75</td>
</tr>
<tr>
<td>C, Calaveras–Hayward–etc. Mean</td>
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<td>56</td>
</tr>
<tr>
<td>−2σ</td>
<td>1.24 × 10¹⁸</td>
<td>6.46 × 10¹⁷</td>
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<td>66</td>
</tr>
<tr>
<td>+2σ</td>
<td>8.97 × 10¹⁷</td>
<td>9.85 × 10¹⁷</td>
<td>1.88 × 10¹⁸</td>
<td>48</td>
</tr>
<tr>
<td>D, northern Calaveras–Green Valley–etc. Mean</td>
<td>6.00 × 10¹⁷</td>
<td>3.78 × 10¹⁷</td>
<td>9.78 × 10¹⁷</td>
<td>61</td>
</tr>
<tr>
<td>−2σ</td>
<td>7.85 × 10¹⁷</td>
<td>1.93 × 10¹⁷</td>
<td>9.78 × 10¹⁷</td>
<td>80</td>
</tr>
<tr>
<td>+2σ</td>
<td>4.37 × 10¹⁷</td>
<td>5.40 × 10¹⁷</td>
<td>9.78 × 10¹⁷</td>
<td>45</td>
</tr>
<tr>
<td>E, Greenville Mean</td>
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<td>1.90 × 10¹⁶</td>
<td>4.97 × 10¹⁶</td>
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</tr>
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<td>+2σ</td>
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<td>All branches of northern San Andreas fault system Mean</td>
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<td>66</td>
</tr>
</tbody>
</table>
data suggest that surface aseismic slip occurs only on faults that exhibit surface creeping behavior interseismically and that earthquakes of greater magnitude (e.g., 1976 $M_w$ 7.5 Guatemala earthquake) appear to have proportionately less postseismic release relative to coseismic slip than do smaller events, such as the 2004 $M_w$ 6.0 Parkfield earthquake, which had no coseismic surface slip and substantial aseismic behavior (Lienkaemper et al., 2006). For the south transition, or Parkfield section (B1, Figs. 6, 8), frequent surface slip rate observation began after the 1966 Parkfield earthquake, initially as aseismic slip and interseismic slip on alignment arrays (Smith and Wyss, 1968; Burford and Harsh, 1980; Lienkaemper and Prescott, 1989). During the 1980s, the USGS installed several additional alignment arrays as part of the Parkfield prediction experiment (Bakun and Lindh, 1985; Baker, 1993), but monitoring of the arrays ended after closure of the prediction window in 1993. Lienkaemper et al. (2006) recovered seven of these arrays following the 2004 Parkfield earthquake. Six arrays on section B1 were used to infer the interseismic creep rates, which are included along with two near-fault GPS station pairs.

Results

The presentation order of the modeling results begins with the main fault branch B, the San Andreas, numerically by section from south to north. Then it continues eastward to branch C (the Calaveras–Hayward–Rodgers Creek–Maacama faults) and branch D (the northern Calaveras–Green Valley–Bartlett Springs faults). Last, the two branches contributing least to seismic release and to overall seismic moment of the NSAFS (Fig. 8), branch A (the San Gregorio fault) and branch E (the Greenville fault) are presented. The primary results are tabulated for all fault sections in Table 1. The observed surface creep rates are compared with the modeled values in Figure 3, mean section depth of creep in Figure 6, and fraction locked in Figure 7. A summary plot (Fig. 8; Table 2) compares the seismic moment release and accumulation rates for the five branches and the NSAFS as a whole.
in the mean surface creep rate used for section B1 (Fig. 3). The Highway 46 array (313 km) exhibits minor creep but lies south of the main 2004 Parkfield rupture that terminated near Gold Hill (307 km), so this minor tail is ignored in the model. Because the Parkfield section represents the southern transition from the rapidly creeping central section (B2) to a fully locked section, surface creep rates decline abruptly here ($\geq 25$ mm/yr to near zero; 275–302 km) in a way that fits poorly near the segment’s north end (Fig. 3, Middle Mtn, XMM4, 282 km). Much more complete and detailed models of coseismic, postseismic, and interseismic slip on the Parkfield section are available in Murray and Langbein (2006). Section B1 is included primarily to develop a comprehensive stress model for the entire NSAFS, especially for the San Francisco Bay area and northward.

Along the San Andreas fault’s central creeping (B2) and north transition (B3) sections, BH80 provided initial creep rates from surveys on several alignment arrays of $\sim 5$–10 yr duration, now supplemented by 2003 surveys on a few of these arrays using a different method by Titus et al. (2006), who found that longer-term rates appear to be distinctly lower. Because BH80 does not supply error estimates, the error is assumed to be $\sim 1$ mm/yr, the error estimated for long-term Parkfield array rates (Lienkaemper et al., 2006). To determine the mean section creep rate for B2, we only used two-thirds of the BH80 rates to give additional weight to some newer, better-determined rates. Figure 3 displays all BH80 creep rates in section B2, including four much higher rates of BH80 that distinctly disagree with all newer long-term data. These higher earlier rates might include transient slip associated with the 1966 $M_w 6.0$ Parkfield earthquake. Alternatively, creep rate at these sites may have simply slowed considerably since the 1970s for unknown reasons.

A significant pulse of transient slip occurred along much of the northern section (B3) of the creeping San Andreas fault following the 1989 $M_w 6.9$ Loma Prieta earthquake (Breckenridge et al., 1997; Nadeau and McEvilly, 2004). However, Titus et al. (2006) found no evidence for any post-2004 Parkfield earthquake transient pulse having reached the fastest-creeping part of the central creeping section (218–266 km; B2). To date, no large transients have been observed on MEE1–MEE2 ($\sim 248$ km), a cGPS station pair installed before the 2004 event.

Despite fairly large uncertainties in observed rates and modeled creep rates for section B2, the modeled uncertainties in depth of creep are relatively low ($\pm 0.7$ km), because
of the high San Andreas driving rate (~34 mm/yr). Similar to the south transition section (B1), the north transition (B3) also fits the data poorly because of its steep gradient in creep rate, declining from >25 mm/yr to near zero in less than 40 km (183–145 km). The mean fit creep rate curve for section B3 is biased slightly to favor the more recent better-determined data (i.e., having lower uncertainties and longer term), whereas the calculated uncertainty range reflects all of the data in Figs. 1 or Table 1. The middle bar indicates the mean of model ranges, and adjacent bars indicate the 2σ uncertainty. The color version of this figure is available only in the electronic edition.

The San Andreas fault north of San Juan Bautista (sections B4 and B5) is locked except for some minor creep (0.5–2 mm/yr) in the southernmost ~10 km, as demonstrated by long-term data from five alignment arrays and now corroborated by a rate derived by a cGPS pair for section B4. Near Point Reyes (section B5), 6 yrs of cGPS data indicate a creep rate of 1.6 ± 0.6 mm/yr (1σ) compared to ~0.1 ± 0.3 (1σ) mm/yr over 27 yrs on the nearby alignment array, SAPR. Although it is possible that creep may be occurring over a broader zone than contained within the 71 m long SAPR, we chose to exclude the currently more uncertain cGPS data from the mean rate used in this model for section B5.

Branch C, Calaveras–Hayward–Rodgers Creek–Maacama–Etc. Fault Sections (C1–C11)

The south end of known creep on fault section C1 is at Pionnee (~191 km, Fig. 2) on what is locally named the San Benito fault (Quaternary Fault and Fold Database for the United States, Data and Resources). Here, the Calaveras fault zone (branch C) lies only 2.7 km from the San Andreas fault (branch B). These two branches, C1 and B3, remain subparallel for the next 30 km until Tres Pinos (site E, Fig. 2) as generally separate, variably dipping structures, based on well-located microseismicity (Watt et al., 2013, 2014). Section C1 is treated as a separate driving source from the nearby San Andreas sections (B2 and B3). Although this simplifying assumption of having the driving slip applied on two deep separate vertical fault patches may or may not be strictly correct, it should produce similar results for the purpose of estimating depth of creep. A greater problem is posed by considerable uncertainty in the four available observed creep rates (sites: CV7S, D, E, and F in Fig. 2a; Table S1) from Hollister (~154 km) southward to Pionnee (site F). Except for site F, the other sites may not completely span the multiple traces that characterize this section of the fault (Quaternary Fault and Fold Database for the United States, Data and Resources). Sites E and F were trilateration sites of short duration and relatively high uncertainty, so they only weakly constrain our knowledge of creep rate. Altogether, it seems appropriate to allow section C1 to include this poorly constrained part of the Calaveras fault, because available surface creep rate data do not allow us to distinguish it from the rest of C1. When modeled as terminating at or near the Pionnee site, there is a steep gradient in creep rate from Pionnee to Hollister, which is consistent with all available data within their large uncertainties (Fig. 3).

North of Hollister, more long-term creep rate data are available for section C1, which span most or all of the fault zone width. A new array (CVCC) was not included in the average rate for C1, because its error is not yet as well constrained as for the other arrays. We did average some cGPS pairs with a similar error, because we are entirely confident that they span the entire fault, which is less certain for CVCC. There is a distinct difference in creep rate estimated for site C of 9.4 ± 0.4 mm/yr (93.2 km, Fig. 3; Grant Ranch trilateration network, ~3 km aperture) and of 12.4 ± 0.5 mm/yr from cGPS pair P218–P241 (106.7 km); both estimates are ±1σ. They are fairly close geographically and both reflect large apertures, so they certainly span the entire fault zone. P218 lies closest to the most recent moderate–large earthquake on this fault, the 2007 M 5.4 Alum Rock earthquake (Murray-Moraleda and Simpson, 2009; Oppenheimer et al., 2010), but its overall displacement signal appears distinctly linear over its entire ~8.5 yr record and thus indicates no significant lasting transient effects from it. The rate difference between Grant Ranch and P218 may indicate an increase in the creep rate gradient between these sites. Both rates are consistent with our simple model within the uncertainties.
The fault traces of section C1 at its north end branch into section D1 through an \( \sim 17^\circ \) releasing bend, while it transfers a greater amount of slip through a compressional left step-over to section C3, the Hayward fault (Fig. 2a and Table 1; Andrews et al., 1993; Simpson et al., 2004). Cluster analysis of GPS velocities (Simpson et al., 2012) clearly illustrates the stronger regional coherence of this Calaveras-to-Hayward slip transfer than compared to the northern Calaveras fault. This motivates the choice of the Hayward fault as the dominant branch and thus the preferred choice for the continuation of branch C. In addition, Figure 3 shows that a greater amount of surface creep transfers from C1 to C3 rather than to D1. Branch C closely follows a small circle path that parallels the plate boundary, a characteristic that it shares with the central creeping section of the San Andreas (section B2). We initially included the southeast extension of the Hayward fault (C2, Fig. 2) but removed it from the model because its long-term slip rate is poorly known and it is far simpler to pose a model with only three intersecting deep dislocations (C1, C3, and D1). No evidence of creep has been demonstrated on section C2.

The Hayward fault (C3, C4, and C5) surface creep observations are the most spatially dense of any fault and have been modeled in detail to infer the depth distribution of creep (Bürgmann et al., 2000; Simpson et al., 2001; Funning, Bürgmann, Ferretti, and Novali, 2007; Lienkaemper et al., 2012; Shirzaei and Bürgmann, 2013). We simplified the Hayward fault to allow comparison to the coarser sections used on most other faults where there are less detailed data. We distinguished the more slowly creeping (i.e., most locked) middle part of the fault as a separate section (C4), from the two faster creeping northern (C5) and southern (C3) sections. Even using such a simple (single creeping patch) method, we find that for sections C3 and C4, which ruptured in the 1868 M 6.8 Hayward fault earthquake (Lienkaemper et al., 2010), the modeled aseismic release rate is exactly consistent with the \( \sim 160 \) yr mean recurrence interval inferred from a two millennia paleoseismic record for similar large earthquakes on section C3 (Lienkaemper et al., 2010, 2012). Section C6 is the continuation of the Hayward fault under San Pablo Bay (Lienkaemper et al., 2012); lacking in observations, its mean surface creep rate is taken to be the average of the rates from sections C5 and C7.

The Rodgers Creek fault (section C7) was long considered a noncreeping fault, but with the installation of the RCSD array (\( \sim 56 \) km) in 2002, a low creep rate (\( \sim 2 \) mm/yr) was detected, prompting us to add the RSCM array (Fig. 2; McFarland et al., 2014). Funning, Bürgmann, Ferretti, Novali, and Fumagalli (2007) inferred creep from permanent scatterer InSAR on C7 north of RCSD of \( 2-6 \) mm/yr, further motivating us to install additional arrays, especially north of RSCM. Currently, the mean creep rate on section C7 is \( 1.5 \pm 0.3 \) mm/yr, based on six alignment arrays and one GPS pair. Overall, section C7 appears to be about \( 90 \pm 10\% \) locked, which still permits significant aseismic release. The location of the maximum creep rate (6.0 \( \pm 0.6 \) mm/yr) estimated by Funning, Bürgmann, Ferretti, Novali, and Fumagalli (2007) from modeling InSAR data is near our array RCMW (\( \sim 64 \) km), which currently averages 3.9 \( \pm 1.4 \) mm/yr and is the highest rate measured on the alignment arrays. However, given the large uncertainty in the RCMW rate, its long-term creep rate could still prove to be as low as the rest of the fault.

The southern Maacama fault (C8) is oriented about 21° counterclockwise to the plate boundary trend (Figs. 1 and 2) and thus might be expected to experience greater fault normal compression than most of the rest of this fault zone; however, fault obliquity is not always a reliable guide to the likelihood of creep occurrence (Bilham and Williams, 1985; Argus and Gordon, 2001). The only alignment array on this section, MSKP (\( \sim 87.6 \) km), has exhibited little if any creep (0.7 \( \pm 0.7 \) mm/yr) in 4 yrs. Apparently, the high heat flow associated with geothermal areas lying to the east is insufficient to encourage creep here, as is found along the central Philippine fault in a geothermal zone (Duquesnoy et al., 1994). Although MSKP spans the known width of the fault zone, it would be advisable to install additional GPS stations west of the fault near the middle of section C8 to corroborate that it is indeed primarily locked.

The central (C9) and northern (C10) Maacama fault sections are separated by a 1–2 km left stepover at Laughlin Ridge (Fig. 2); however, creep rates are roughly similar on both sections. The 19 yr Ukiah array (MUKI, \( \sim 144.5 \) km) rate of 4.3 \( \pm 0.8 \) mm/yr has nearly the same mean rate as that derived from averaging multiple station pairs of mostly campaign GPS over 4–7 yrs (4.2 \( \pm 0.2 \) mm/yr, 1 standard error of mean of N used [SEM]). The 20 yr Willits array (MWIL, \( \sim 179.1 \) km) creep rate of 5.7 \( \pm 0.1 \) may only be compared with recent (\( \sim 2 \) yr) campaign GPS data spanning a much broader, \( \sim 2-3 \) km aperture. The preliminary average creep rate from the GPS data on section C10 is \( \sim 9 \) mm/yr, which equals the assumed loading rate and thus would indicate no locking. The GPS data are excluded from the mean rate for the C10 section used in the model, although if their current mean proves correct over a longer period, these GPS rates would suggest much greater aseismic release occurs here than is observed within the \( \sim 125 \) m long MWIL array. Creep rate on MWIL has been more variable over time than many arrays, averaging about 8 mm/yr for its initial \( \sim 3.5 \) yrs, but over the past 7 yrs it has averaged only 2.8 mm/yr. The cause of this variability is unknown, thus using only the one available long-term average seems appropriate here. We have concerns that the high short-term (\( \pm 2 \) yr) creep rates estimated from the GPS data might include a slip transient, which can occur occasionally along creeping faults and can be significantly higher than the long-term creep rate; for example, \( \sim 2 \) cm of completely aseismic slip occurred in 1996 along a 6 km extent of the southern Hayward fault in only two days (local long-term creep rate \( \sim 0.9 \) cm/yr; Lienkaemper et al., 2012). Caution is advised when using the model results for section C10, because the formal multiple linear regression error for MWIL also does not reflect larger uncertainties inherent in having no creep rate data for the rest of...
C10 to the north of the Willits area. The northern part of C10 is an extension of the mapped Maacama fault near Laytonville that is assumed to connect with the Brushy Mountain shear zone, thence to the south end of the Brichelnd fault, but detailed mapping of active traces is lacking north of Laytonville.

Murray et al. (2014) model creep on the San Andreas, Maacama, and Bartlett Springs faults exclusively using GPS data (i.e., no alignment array surface creep rates) to constrain creep rate along these faults. Their inverse modeling method simultaneously solves for driving depth and driving rate. This differs from our choice of consensus values of these parameters, which depend primarily on depth of the base of microseismic activity to infer driving depths and long-term geologic slip rates to infer driving rates. The geologic rate is poorly constrained for the Maacama (Prentice et al., 2014) and unknown for the Bartlett Springs fault, so consensus slip rates assume continuity of geologic rates observed farther south. Qualitatively, their model finds similar ranges of surface creep rate for the Maacama fault, except for much higher values near Willits, because it uses only the short-term (≤2 yr) high GPS rate in Willits. Their model suggests a significantly higher creep rate on the southern Maacama near where our alignment array detects almost none (0.7 ± 0.7 mm/yr); however, there are no GPS stations used west of the fault to constrain this section. Unlike our model, no locking patch is assumed in their inversion, and no fully locked areas are found in the lower layer of their two-layer model. Given considerable differences in driving rates and depths between our models, it would probably be best to compare our results in terms of moment accumulation. Murray et al. (2014, their table 1) estimate that moment sufficient for an $M_w \sim 7.0$–7.1 earthquake has accumulated over the past ~500 yrs. Our model permits accumulation over that period of $M_w \sim 7.2$ on our section C10, which is 91 km long, half of which lies north of their modeled Maacama fault. Thus our deficit rates appear to be approximately consistent for this section of the Maacama fault.

We continue section C11 along the Briceland fault and a little into the Bear River fault to near Petrolia. Because sections C11 and D10 are near the Mendocino triple junction (Fig. 1), it is especially important that we estimate creep rate from GPS velocity using only the Sierra–Pacific plate parallel velocity components for both observed and calculated rates (this is done for all GPS pairs) to exclude any convergent deformation caused by the subducting slab (Williams et al., 2006). A pair of cGPS velocities (P163–P158) suggests a substantial creep rate (4.0 ± 0.4 mm/yr) may occur nearly as far north as Honeydew, California, but it remains unknown if any creep continues on fault branch C north of the downgoing Gorda plate slab (~ −300 km, Figs. 1, 2). Mapping of active fault traces in this area is needed to understand in detail exactly what structures may be involved in this apparent northward continuation of surface creep.

Branch D, Northern Calaveras–Green Valley–Bartlett Springs–Etc. Fault Sections (D1–D10)

Branch D makes an ~17°-clockwise (extensional) bend away from the north end of section C1, whereas branch C continues northward along a trend maintaining much better alignment with plate boundary motion. Beginning as the northern Calaveras fault (D1–D2); it then steps right ~5 km to the Green Valley fault zone (GVF, D3–D5), as described in Lienkaemper, Baldwin et al. (2013), composed of the southern GVF (D3, includes Concord fault) and northern GVF (D4–D5). Within section D5 (Hunting Creek fault), the GVF bends back, ~17° counterclockwise, into the Bartlett Springs fault zone (BSF, D5–D9; Lienkaemper and Brown, 2009; Lienkaemper, 2010), to a more nearly plate-boundary aligned trend.

The northern Calaveras fault (D1–D2, Figs. 1 and 2) comprises a shorter (14 km), faster (~4 mm/yr), steadily creeping section (D1) near where it branches from the Calaveras fault (C1) and a longer (27 km), slower (~2 mm/yr), episodically creeping section (D2, Fig. 3). Episodic creep is the occurrence of distinct creep events, which may last seconds to weeks and tend to be large relative to the long-term creep rate of a site. Such creep events may range in amplitude from a few millimeters up to about 2 cm. Creep release on the slower section (D2) has had an unusually episodic character compared with other faults in the region. All three arrays on D2 exhibited lengthy periods entirely without any creep (varying from 3 yrs [CVSP] to 12 yrs [CVCP; McFarland et al., 2014]) with no recognized cause for this behavior. Similar highly episodic behavior has been identified at many other sites along the rest of branch D. Such episodic behavior seems to be more prevalent for sites with relatively low creep rates (e.g., <4 mm/yr), thus it can require many years and (rarely) decades to establish reliable long-term creep rates on slowly creeping faults. Thus, the lack of long-term creep data is a major cause of concern for the more remote northern sections, D4–D10, where currently sample durations are only 2–7 yrs. Nevertheless, the average rate for section D2 now appears to be sufficiently well determined that it appears to be mostly locked, whereas section D1 appears to be mostly creeping.

The southern Green Valley fault (D3), which herein includes the Concord fault, is separated from section D2 by an ~5 km right (extensional) stepover. There are now abundant long-term creep rate observations on D3 that average ~3 mm/yr, which the model mostly fits well (Fig. 3). However, at the north end of D3, creep seems to diminish more abruptly than modeled, at array GVCL (~39.6 km; raw 1.3 ± 1.3 mm/yr, or rainfall-corrected 1.3 ± 0.3 mm/yr 2007–2013) and GPS pair LB02–LB01 (~38.7 km; 0.8 ± 0.7 mm/yr). If these low rates at the north end of section D3 are confirmed with additional data, then an additional transitional section could be added to the model to better approximate a reduction of creep approaching the Berryessa fault section, D4, perhaps because D4 is locked.

A very preliminary mean creep rate of ~0.3 ± 1.3 mm/yr on section D4, the Berryessa fault
(-56.1 km; 2010–2013) allows with large uncertainty that this section may be entirely locked (Fig. 3). Section D4 is only 23 km long and is much more complex than indicated in Figure 2b, which only shows the eastern trace to simplify the modeling. Its two active traces together form an ~2 km left (compressive) stepover (Lienkaemper, 2012a,b) that may be the cause of the apparent locking. So far no creep has been observed on the western trace either (McFarland et al., 2014). The Hunting Creek fault (D5), the transitional section from the GVF to BSF, exhibits episodic creep on array HCHC (~78.6 km) that currently averages 1.6 ± 0.6 mm/yr over a 6 yr period. This rate is similar to the nearby GPS pair rate of 2.0 ± 0.7 mm/yr (~75.0 km, HC04–HC03). The consistency between the GPS and alignment array data suggest that locking may slightly exceed aseismic release on D5, but uncertainties remain somewhat larger here than for most urban fault sections to the south.

Near Highway 20, the next short (15 km) BSF section (D6) has a 3 yr creep rate of only 1.2 ± 0.7 mm/yr at array BS20 (~102.1 km) but a much higher 4 yr rate of 4.4 ± 0.7 mm/yr from a nearby GPS pair 3751–HP11 (~102.6 km). Because of the considerable difference in the surface creep rate estimates from the GPS pair and the alignment array, the creep depth and locked fraction remain poorly constrained but suggest aseismic release may dominate here. The array site might be dominantly episodic and simply not had any sizeable creep event expressed on the alignment array yet, or perhaps some of the deformation zone falls outside of the array.

Near Bartlett Springs, BSF section D7 has surface creep rate estimates derived from three GPS pairs (Tables 1 and 5; S2) that average −0.2 ± 0.7 mm/yr (1-SEM). Earlier results from the BSNW alignment array are not usable because of an unstable monument, which was relocated to a stable position in 2013. D7 is the only section of either GVF or BSF that currently is estimated to exhibit dominantly locked behavior.

The Lake Pillsbury section (D8) of the BSF has a 7 yr creep rate of 3.1 ± 0.2 mm/yr (BSLP, −165.7 km) that is now well determined despite its episodic behavior. However, 9 km north of BSLP, a 3 yr rate of 7.0 ± 0.7 mm/yr from GPS pair WOTI–ISLE (~174.4 km) indicates the section average must exceed the BSLP rate. A minimum mean surface creep rate of ≥4.4 ± 0.5 mm/yr for section D9 is calculated as the average of the BSLP rate of 3.1 ± 0.2 mm/yr and the WOTI–ISLE minimum value of ≥5.6 for (7.0 less 2σ). Hence, section D8 appears to be dominantly releasing aseismically, but its mean creep rate and hence its aseismic release rate remains poorly constrained for hazard purposes.

Comparing our model results with those of Murray et al. (2014) for the BSF sections that are included in both of our models (D7 and parts of D6 and D8), we find generally similar results of roughly similar creep in the near surface for sections D6 and D8, and greater locking in section D7. However, their model places rates of aseismic slip in their lower layer, comparable to their driving rate, thus suggesting that no significant seismic moment accumulation occurs at depths of 5–13 km. Even the lower layer below D7 models in such a way as to not accumulate significant moment, despite our models both agreeing on the absence of surface creep here. A comparable result is not obtainable with our approach, because our model assumes a lower fully locked layer that is overlain by a creeping layer. They note that in any “…inversion approach the ability of geodetic data to resolve spatially variable slip at depth on a near-vertical fault is limited.” King and Wesnousky (2007) suggest a possible solution to this dilemma, their analysis of background seismicity in NSAFS shows seismic slip dominantly occurring at depths between 4 and 12 km with its maximum at about 8 km, and they propose that greater aseismic release may be occurring above and below these depths. Thus, a possible alternative to our two-layer locking model approach and that of Murray et al. (2014), might be a three-layer model that requires a locked layer of variable thickness, perhaps centered near 8 km, and sandwiched between an upper creeping zone and a lower creeping zone. Such an approach may produce an intermediate solution between our results and those of Murray et al. (2014) that perhaps could result in somewhat more realistically located locked areas.

The Round Valley section (D9) of the BSF is separated from section D8 by an ~2 km left (compressional) stepover (Lienkaemper and Brown, 2009; Lienkaemper, 2010). Near the south end of D9, the rate at BSRV of 0.4 ± 1.5 mm/yr indicates no significant creep (2008–2011), although an adjacent wooden fence line of unknown age (but a few decades by appearance) is dextrally offset at the fault by 140 ± 48 mm (1σ), calculated using multiple linear regression. The GPS pair nearest our arrays (P319–CVR3, −205.6 km) yields a low, sinistral surface creep rate of −1.8 ± 0.6 mm/yr (2001–2013), which suggests that creep rates at the south end of section D9 may be much lower, although irregularity in the P319 velocity record indicates that this cGPS station might be somewhat unstable. Because of its sinistral sense, we exclude the P319 creep estimate in calculating the mean section rate. Because these lower alignment array rates are observed closer to a 2 km restraining left stepover (~200 km), the surface creep may occur more sporadically, perhaps comparable to the episodically creeping sites of section D2, that is, characterized by long periods without creep followed by either single creep events or a few years of steady creep.

The central part of section D9 appears to exhibit creep based on both older trilateration and recent GPS work. Lisowski and Prescott (1989) estimated a high creep rate (~8 mm/yr) based on measurements made from 1985.164 to 1989.381 (decimal years) on the Cove1o-Poonkinkey trilateration line, which crosses the middle of section D9 at a low angle (USGS) and was based on a single repeat measurement. Using archived data (USGS), we compute a 1989 maximum creep rate from this measurement of 8.6 ± 0.6 mm/yr using the formal uncertainty archived with these data. It is a maximum, because it does not compensate for some unknown but fairly minor strain component. A nearby GPS pair P321–CVR3 (~228.5 km) that samples
the same middle part of section D9 yields a surface creep rate of 6.4 ± 0.4 mm/yr rate for 2001–2013. The original 1989 Covelo benchmark is no longer repeatable (J. Svarc, oral comm., 2013) The high 1989 rate may include a transient signal, while both stations P321 (continuous) and CVR3 (survey) have shown steady linear rates of movement over a 12 yr period. Thus, we consider the current GPS-derived creep rate to be more reliable here and did not use the older trilateration rate. Nevertheless, the lack of well-demonstrated surface creep across our arrays near the south end of D2 remains problematic, suggesting that creep may be either episodic or else it may occur on structures other than the mapped fault traces crossed by our arrays.

In summary, the three GPS pairs across or near section D9 average 4.7 ± 2.3 mm/yr, but the mean of all of the GPS and the two alignment array observations taken together is only 3.3 ± 1.7 mm/yr. Altogether, current results seem somewhat equivocal. Some significant aseismic release seems likely, at least for the middle of this section, but how much occurs overall still remains highly uncertain.

North of Round Valley, the BSF makes a pronounced (~24°) releasing double bend through the Lake Mountain fault zone of Herd (1978), where to the north it resumes trending subparallel to the plate boundary along the Eaton’s Rough fault. These two fault zones are modeled together as section D10. Two GPS pairs together allow a low dextral surface creep rate of 1.9 ± 0.4 mm/yr modeled on section D10 well north of the southern edge of the subducting Gorda plate. An ~3.5 km dextral bedrock offset along the Eaton’s Rough fault was recognized by Kelsey and Carver (1988) near the southern GPS pair P168–P326. The northern pair P169–P326 lies near the transition of the Eaton’s Rough fault into a complex zone of mixed thrust and dextral faulting (Kelsey and Carver, 1988).

Branch A, San Gregorio Fault (A1), and Branch E, Greenville Fault Sections (E1–E3)

The San Gregorio fault (A1) branches southward from the San Andreas fault (B4 and B5) near Bolinas Lagoon (Fig. 2; ~6 km) and lies mainly under the Pacific Ocean, so few geodetic measurements are currently possible. Array SGWP (40.4 km) indicates that no creep (~0.2 ± 0.04 mm/yr) occurs on A1, although the other array SGPR (69.4 km) has a calculated rate of 1.0 ± 0.1 mm/yr on an ~450 m long array, which indicates minor creep may be possible. However, the latter array’s great length and unusually noisy data also make it plausible that there is no creep. Lack of creep on A1 is also supported by the GPS pair P277–P534 (75.1 km) located near SGPR, which yields a much lower rate of 0.1 ± 0.5. This GPS-derived surface creep rate is the primary basis of the model of A1, that is, it allows near-zero creep within the uncertainty range allowed by the GPS pair, and it is strongly corroborated by the other section A1 array SGWP. For completeness of the regional model, we include an extended San Gregorio fault source in the model south of A1 with the same driving rate, but we do not model creep south of A1.

The Greenville fault (E1–E3) may not strictly branch from the NSAFS. Its closest approach is to branch D, where section E3 steps left (compressionally) ~8–10 km through Mt. Diablo to the Concord fault in section D3 (Unruh and Sawyer, 1995) with a large (~14 km) gap. Lienkaemper, Barry, et al. (2013) recently modeled Greenville fault creep with the same observations used in the current model, but they used a 2D approach and the equations of Savage and Lisowski (1993) rather than those of Okada (1992). The northern section E3 has a 47 yr surface creep rate of 2.0 ± 0.3 mm/yr from the Green trilateration short-range network (59.2 km) is corroborated by array GALT (58.4 km) as 1.7 ± 0.4 mm/yr, and GPS pair DIAB-P230 modeled creep rate of 2.7 ± 0.2 mm/yr. The current 3D model estimates E3 to be predominantly creeping, with only ~15% (3%–31%) locked, and that creep extends much deeper (to ~14 ± 2 km), rather than the ~6 (3–11) km estimated using the 2D model. However, results for the southern sections (E1–E2), based only on creep rates inferred from GPS pairs remain similar for both the 2D and 3D models, finding both sections dominantly locked. The significantly greater creep releases on section E3 in the 3D model, makes a sizable difference in total moment accumulation expected for the entire fault. Using the 2D model, a full rupture of the locked areas can produce an $M_w$ 6.9 earthquake (using the magnitude–area [MA] relation of Hanks and Bakun, 2002, 2008), but the 3D model produces only $M_w$ 6.7, capable of loading in ~400 yr rather than ~600 yr for the larger event. The estimates of surface creep rate for the southern Greenville fault (E1–E2) are still based only on GPS station pairs located farther from the fault than most of those included in this analysis, so these model results need to be corroborated by smaller aperture observations, such as an alignment array or closer GPS stations. We have recently installed four additional alignment arrays along the Greenville fault and are now monitoring for creep rate on all of its sections.

Discussion

Rates of Seismic Moment Accumulation and Aseismic Release by Fault Branch and Systemwide

For each of the five branches of the NSAFS, we now compare the proportion of seismic moment that releases aseismically with the proportion accumulating by locking or by pinning (Fig. 8; Table 2). The model is intentionally coarse, having divided an ~2000 km extent of faults into only 29 sections, and by using only the average of each section’s long-term surface creep rate to infer its depth of creep. Such a coarse fault model may thus miss considerable detail in spatial and temporal variations. However, for the majority of sections, long-term creep rate data tend to be sparse, thus averaging them provides a more robust way to characterize surface creep, except where creep rate gradients are unusually steep (e.g., section B1), and the overall results of this...
model are similar to much more detailed modeling of sections where data are plentiful, such as the Hayward fault (C3, C4, and C5; Lienkaemper et al., 2012). We estimate that the entire northern San Andreas fault system releases only ~28% aseismically, primarily because the dominant fault (the San Andreas) releases only ~21% aseismically. The two next largest branches C (Calaveras–Hayward–etc) and D (northern Calaveras–GVF–BSF–etc) each release ~40% aseismically, compared to only ~20% on branch B (San Andreas). The vast majority of aseismic moment release (84%) by the NSAFS occurs within ≤8 km of one small circle path along the Sierra–Pacific plate boundary (near the central grid line of Figs. 1, 2; Argus and Gordon, 2001). This higher proportion of aseismic release starts along the creeping San Andreas (B1, B2, and B3) and continues along branch C, from the southern Calaveras (C1) as far north as the Mendocino triple junction (C11). In contrast, significant surface creep release does not occur along those sections of the San Andreas located between double restraining bends, such as through the Santa Cruz Mountains (100–150 km, Fig. 1) or the Mojave section in southern California (Argus and Gordon, 2001). However, north of the Santa Cruz Mountains, much of sections B4 and B5 aligns reasonably subparallel to the plate boundary but does not exhibit any recognizable creep. The southern sections of branch D (D1–D5) comprise an extensional double bend, which should encourage creeping behavior by relaxing fault normal stress, and these sections are overall dominantly creeping. Although the northern sections (D5–D10) mostly trend subparallel to the plate boundary, they also appear to accommodate creep about as well, especially in two sections (D5 and D8), but creeping behavior in other sections is still poorly characterized. Creep along the NSAFS generally appears to be favored where faulting occurs along a relatively smooth plate boundary parallel path, but other factors, including variation in fault zone mineralogy (Moore and Lockner, 2013), frictional strength (Morrow et al., 2010; Lienkaemper et al., 2012), and heat flow (Duquesnoy et al., 1994), may also contribute to creeping behavior. Conversely, local fault complexity may inhibit creep, particularly where it increases the normal stress. For example, the Berryessa fault (D4) includes an ~2 km compressional stepover that preliminary data suggest may suppress creep. Similarly, at the south end of Round Valley proximity to an ~2 km constraining stepover in the Bartlett Springs fault also appears to reduce creep. We agree with Argus and Gordon (2001) that, at best, the degree and sense of obliquity of any particular fault section to the direction of plate motion can only partially explain the likelihood of the occurrence creep and that many other factors can have equal or greater importance.

Caveats Regarding Use of Seismogenic Scaling or Aseismicity Factors

A quantity such as fraction locked is a necessary component of seismic-hazard assessment that can be used to reduce effective source areas for modeling of earthquake size and frequency, but it should be used with considerable caution. For example, we use a low magnitude cutoff (M ≥ 5.5) for purposes of eliminating the effects of moderate-to-large earthquakes in the calculation of the long-term, interseismic surface creep rates used in the model. For most fault sections, this probably poses no conflict with seismic hazard calculation; however, for faster creeping sections (especially for section C1, which has historically produced at least two moderate-to-large earthquakes [M 5.8–6.2, Oppenheimer et al., 1990]), this becomes an important consideration. Current seismic-hazard analysis in California considers M ≥ 6.7 (Working Group on California Earthquake Probabilities, 2003, 2008; Field et al., 2013) as large earthquakes, although lower magnitudes are involved in the modeling. To evaluate the fraction locked specifically for earthquakes of M ≥ 6.7, the total amount of seismic moment release accompanying M 5.5–6.7 earthquakes first needs to be accounted for separately, including coseismic slip, afterslip, and propagation of long-lasting transients such as those observed at array CVCR. The creeping San Andreas sections (B1, B2, and B3) also require some deduction for any similar moment release associated with M < 6.7 events. Other than the Parkfield section (B1) that has had frequent well-documented large–moderate (M ∼ 6) earthquakes, the other two sections (B2 and B3) possibly experienced sizeable moment release associated with earthquakes in the M 5.5–6.7 range during the nineteenth century (Toppozada et al., 2002), but their locations and sizes are highly uncertain.

In contrast to the southern and central Calaveras fault section (C1), for the northern Calaveras (D1 and D2) we estimate a locked fraction of ~66%, and it has not yet exhibited detectable changes in creep rate in response to historical seismicity during the instrumental period. However, an M ∼ 5.8 earthquake in 1861 possibly ruptured a few kilometers near the north end of section D2 (Bakun, 1999), where it currently exhibits a very low creep rate (1–2 mm/yr).

It is especially important for creeping faults to obtain fault-specific information about their paleoearthquake histories, because seismic-hazard evaluations usually depend on MA relations (e.g., Hanks and Bakun, 2002, 2008) to estimate magnitude for a proposed earthquake source area. However, these MA relations are based primarily on noncreeping faults and thus need to be used with great caution for faults that exhibit a sizeable aseismic release rate. For example, Hayward fault sections C3–C4 ruptured in 1868, producing an M ∼ 6.8 earthquake (Bakun, 1999). On section C3, the mean recurrence interval over the past two millennia is estimated at ~160 yrs (Lienkaemper et al., 2010). However, using the fraction locked from the model to adjust the equivalent rupture area for C3–C4, we obtain only M 6.6 from the Hanks and Bakun (2002, 2008) MA relation, which would be capable of reloading in only 80 yrs. The observed longer Hayward fault recurrence is apparently associated with a high friction patch in section C3 (Lienkaemper et al., 2012). This example makes it clear that for creeping faults one cannot necessarily rely on generic MA
relations to correctly characterize their likely magnitude-frequency distributions. It is especially important for creeping faults to consider historical and prehistoric rupture histories and knowledge of any episodes of transient aseismic release and afterslip behavior.

Similarly, the southern Green Valley fault (D3) has a millennial mean recurrence interval based on its paleoearthquake record, estimated at $\geq 250$ yr when its long (~400 yr) open interval or time since previous large earthquake is included (Lienkaemper, Baldwin, et al., 2013). Application of the calculated fraction locked for D3 (57%) to its area using the Hanks and Bakun (2002, 2008) relation estimates an $M \geq 6.8$ earthquake that can reload in only ~140 yr. Unlike the approximately Gaussian or normal recurrence behavior as found for the Hayward fault, the southern Green Valley fault (D3) probably can produce longer (higher magnitude) ruptures, including the northern Green Valley sections (D4 and D5) that can greatly increase some recurrence intervals. Separated from section D3 by an ~2.7 km extensional stepover, the Berryessa section (D4) appears to be almost locked. This section (D4) last ruptured approximately at the same time as the previous southern Green Valley section (D3) rupture (Lienkaemper, 2012a,b), so the most recent earthquake on these two sections (D3–D4) may have ruptured through the extensional stepover and thus many may have been considerably larger (up to $M \geq 7.1$) than a single-section rupture of the southern Green Valley fault (D3). As a result, the potential for connectivity between sections separated by moderate size (1–4 km) geometric discontinuities can be another important factor in estimating sizes of future ruptures (Wesnousky, 2008). The choice of section boundaries for this model is not intended as indicative of potential seismic source extent, but was chosen specifically for purposes of modeling creep extent with available, often-sparse data partly based on fault geometry and also on any known distinct changes in creep rate along strike. Some of the longest sections have included historical short ruptures of varying lengths (e.g., section C1) within them, and many shorter sections may be capable of linking with adjacent sections to form longer ruptures.

Another caveat of the model is that it does not account for uncertainty in either the down-dip width of faulting or the driving slip rates. These uncertainties should have a second-order effect of increasing the uncertainties in depth of creep and fraction locked but do not affect the mean values. \cite{footnotes} See the electronic supplement describing the sensitivity to fraction locked to the uncertainties in driving rate and driving depth. Driving rate uncertainty appears to have a much greater impact than driving depth uncertainty. Our primary focus has been to test the direct effect of the uncertainties in the surface creep rate observations to help evaluate the effectiveness of the creep-monitoring program. Furthermore, we believe that the surface creep rate observations should provide a strong constraint of fraction locked for purposes of seismic-hazard assessment, which includes uncertainties in the driving rates and driving depths separately. See electronic supplement for sensitivity to uncertainty in the driving parameters. Also of concern is that some sections include only one or very few creep observations in estimating the mean rate; thus, in such cases, the formal errors may seriously underestimate real uncertainty. For example, one section (C10) has only one well-constrained rate representing a 91 km long fault. Clearly, more data are needed to fill the largest remaining gaps in near-fault creep monitoring.

As a final caveat, all modeling approaches have biases and uncertainties resulting from their underlying assumptions. In the presentation of the Bartlett Springs fault results (following discussion of sections D6–D8), we contrast the results of the Murray et al. (2014) model for these sections to ours. We note that neither model is designed to locate realistic asperities in the depth ranges where they actually tend to occur in the NSAFS. Analysis of King and Wesnousky (2007) found slip in NSAFS seismicity is greatest between depths of 4 and 12 km, with the greatest slip near ~8 km depth, and they have suggested patches of stable sliding or creep lie above and below this region of greatest slip. Our model forces locking to occur at the base of the seismogenic zone, whereas the approach of Murray et al. (2014) may tend to find no locking near the base of the seismogenic layer, where their model is most undetermined. Using a three-layer model similar to the conceptual model proposed by King and Wesnousky (2007) might produce locking zones that may be more physically realistic than our simple model despite the sparseness of available data. In the future, models that account more realistically for the dynamic frictional properties of the creeping areas (e.g., Barall et al., 2008; Lozos, 2013) may do a better job of evaluating the seismic potential of creeping faults, provided asperities also are mapped reasonably well.

Urban Creeping Faults with Paleoseismic Constraints

The extent of a creeping fault’s largest locked areas, along with its slip rate, directly relates to both the expected magnitude and frequency of its ruptures (Lienkaemper et al., 2012). Our results indicate that three urban creeping faults have large locked areas that have not ruptured in a major earthquake ($M \geq 6.7$) in historical times: Rodgers Creek (C7), northern Calaveras (D1–D2), and southern Green Valley (D3). In addition, the southern Hayward fault (C3–C4) already produced an $M \geq 6.8$ event in 1868, but it is now approaching its mean recurrence time based on paleoseismic evidence (Lienkaemper et al., 2010). Table 3 summarizes paleoseismic estimates of the age of the most recent earthquakes and mean recurrence intervals for significant, groundrupturing earthquakes. Using the mean rate of accumulation of seismic moment calculated from the model for each of these faults and the time elapsed since their most recent earthquake, we estimate that all four of these faults have accumulated sufficient moment to produce a major earthquake. In addition, three faults appear to be nearing or have exceeded their mean recurrence time: the Hayward, Rodgers Creek, and Green Valley.
Summary and Conclusions

Over the past decade, knowledge of the extent of creep along the five branches of the NSAFS has advanced considerably, due to the rapid expansion of the alignment array and GPS station networks. In the urban San Francisco Bay area, few large gaps in monitoring remain. Two large gaps remain in the remote north Coast Range (sections C10, C11, and D10) and uncertainties along branch D, the northern Green Valley and Bartlett Springs faults remain large, mainly because of the need for longer monitoring periods.

The model used to estimate average depth of creep and fraction locked provides a uniform means to compare results along the entire NSAFS, although more detailed models are available for a few sections. The modeled values of the fraction locked give a first-order estimate of a factor that may be used for seismic-hazard modeling of $M \geq 6.7$ earthquakes, provided some caveats are observed. For predominately creeping sections of the central San Andreas and Calaveras faults (B1, B2, B3, and C1) one must first account for all modes of seismic moment release associated with $M < 6.7$ earthquakes, including large, prolonged aftershock and other forms of transient postseismic release that might be expected from moderate-to-large earthquakes occurring on predominantly creeping faults. Lack of sufficient long-term creep data and uncertain historical earthquake records in these sections makes modeling of seismic hazard for such faults highly problematic. Seismic-hazard estimates from creeping faults may be represented poorly by existing empirical magnitude–area relations. Therefore, it can be especially important for creeping faults to obtain fault-specific paleoearthquake histories and to realistically characterize the potential for multisecton connectivity.

For major creeping faults in the urban San Francisco Bay area where paleoseismic information is available (Table 3), using these improved estimates of the fraction locked provides a more accurate assessment of the current potential for major ($M \geq 6.7$) earthquakes, based on the amount of seismic moment that they may have already accumulated. Using the modeled mean accumulation rates, four urban creeping faults appear to have already accumulated enough seismic moment to produce major earthquakes: Hayward ($M \geq 6.8$), Rodgers Creek ($M \geq 7.1$), northern Calaveras ($M \geq 6.8$), and Green Valley ($M \geq 7.1$).

Data and Resources

All data used in this paper came from the sources listed below and in the electronic supplement. Continuous and survey-mode GPS velocity data are from the U.S. Geological Survey: (1) USGS, GPS velocities for San Francisco Bay area, (http://earthquake.usgs.gov/monitoring/gps/); last accessed May 2013, except branch E data downloaded in 2012; sections north of San Pablo Bay last accessed July 2014) and (2) USGS, geodolite trilateration data set (http://earthquake.usgs.gov/monitoring/deformation/geodolite/); last accessed November 2013).


Acknowledgments

U.S. Geological Survey National Earthquake Hazard Reduction Program funded this investigation: 9939-0KR02 (USGS) and G10AC00139 (San Francisco State University [SFSU])! Many thanks to dozens of SFSU student researchers, who since 1979 have conscientiously operated theodolites and precisely set targets, resulting in over three decades of accurate creep measurements. Special thanks to Theresa Hoyt, the longest serving Creeper, who annually verifies all calculations prior to formal archiving. We appreciate the assistance of many private landowners who allow us

### Table 3

<table>
<thead>
<tr>
<th>Fault</th>
<th>Section</th>
<th>Mean</th>
<th>Recurrence Interval (yr)</th>
<th>Locking Moment Rate, Mean (N·m/yr)</th>
<th>Locking Area, Mean (km²)</th>
<th>$M_w^*$</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hayward, south</td>
<td>C3, C4</td>
<td>1868</td>
<td>145</td>
<td>1.14 x 10^17</td>
<td>423</td>
<td>6.8</td>
<td>Lienkaemper et al. (2010)</td>
</tr>
<tr>
<td>Rodgers Creek</td>
<td>C7</td>
<td>1745</td>
<td>268</td>
<td>1.88 x 10^17</td>
<td>697</td>
<td>7.1</td>
<td>Hecker et al. (2005)</td>
</tr>
<tr>
<td>Northern Calaveras</td>
<td>D1, D2</td>
<td>1740</td>
<td>273</td>
<td>6.39 x 10^16</td>
<td>355</td>
<td>6.8</td>
<td>Schwartz et al. (2014), Kelson et al. (2008), Weldon et al. (2013)</td>
</tr>
<tr>
<td>Green Valley, south</td>
<td>D3</td>
<td>1609</td>
<td>404</td>
<td>1.06 x 10^17</td>
<td>592</td>
<td>7.1</td>
<td>Lienkaemper, Baldwin et al. (2013)</td>
</tr>
<tr>
<td>Berryessa</td>
<td>D4</td>
<td>1623</td>
<td>390</td>
<td>5.73 x 10^16</td>
<td>319</td>
<td>6.9</td>
<td>Lienkaemper (2012b)</td>
</tr>
</tbody>
</table>

*Mean magnitude from seismic moment accumulated on mean modeled locking area since time of mean most recent earthquake age; unit-modified Hanks and Kanamori (1979): $M_w = (2/3)\log_{10}(M_0 \times 10^{-10}) - 10.7$(N-m).

1Using Brownian passage time curve fit to paleoearthquake record (i.e., recurrence distribution is non-Gaussian).
access to perform surveys. Reviews by Diane Moore, Ben Brooks, and two anonymous reviewers greatly improved the manuscript.

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U.S. Geological Survey 977
Merlo Park, California 94025
jlienke@usgs.gov
simmop@usgs.gov
(J.J.L., R.W.S.)

Geosciences Department
San Francisco State University
San Francisco, California 94132
forrestmcf@hotmail.com
caskey@sfu.edu
(F.S.M., S.J.C.)

Manuscript received 25 April 2014.