Results for aseismic creep on the Hayward fault using polarization persistent scatterer InSAR

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\textbf{Abstract}
We present new results for shallow aseismic creep on the Hayward fault in California using a new DInSAR technique. This method not only provides, for the first time, the ability to map the displacement field on both sides of the fault, it does so over a much shorter time period than earlier results. The results provide a good match in the near-field to both the regional continuous GPS velocities and data from an alinement network that measures long-term creep along the fault. The average slip rate for the northern segment of the Hayward fault is \(\sim 4.4\) mm/yr between 2008 and 2011, slightly less than that estimated for longer time periods, suggesting that the slip rate may not be constant. If the slip rate along the fault is variable on the decadal or longer scale, current estimates of its earthquake potential and the associated hazard associated with the slip rate deficit may need to be revised from previous estimates. We demonstrate the potential impact of this method to better define the spatial and temporal complexity of aseismic slip and estimate the accumulated elastic strain along one of the most significant sources of seismic hazard in the San Francisco Bay area.

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\section{1. Introduction}

Ground deformation related to the relative tectonic motion between the North American and Pacific plates (\(\sim 39\) mm/yr) is accommodated along a series of well-developed sub-parallel, primarily strike-slip faults in northern California including, but not limited to, the San Andreas, Hayward and Rodgers Creek faults (Fig. 1a) (Argus and Gordon, 2001; dAlessio et al., 2005). Areas of moderate-to-high topography occur as the San Andreas fault passes through the San Francisco Bay Area. Local contraction is accommodated through the Santa Cruz mountains and stepovers between the Greenville and Concord faults to the east and the Hayward and Calaveras faults to the south (Argus and Gordon, 2001). In the last decade, estimates suggest that as much as 60\% of the relative motion is accommodated by the San Andreas fault alone (dAlessio et al., 2005; Jolivet et al., 2009), so that almost half of the strain accumulation is apportioned onto secondary faults (Fig. 1b). In particular, the Hayward fault, a right-lateral structure that parallels the San Andreas for nearly 100 km, accommodates much of the remaining plate-boundary strain in the San Francisco Bay Area. The Hayward fault can be divided into two segments: a southern and a northern segment separated by a locked region between Oakland and Fremont (Fig. 1a) (Schmidt et al., 2005).

In recent years the Hayward fault has received significant attention from the seismological community, in large part because of the seismic hazard associated with its historic seismicity. While the northern segment may not have ruptured in approximately 300 yr, the southern segment was responsible for the M\(\sim 6.8\), 1868 earthquake (Yu and Segall, 1996; Bakun, 1999; Bürgmann et al., 2000; Schmidt et al., 2005). Paleoseismic records suggest that large events similar to the 1868 event occur with a mean recurrence interval of 161 \(\pm 65\) (1\(\sigma\)) yr (Lienkaemper et al., 2010). In addition, its complex behavior includes aseismic creep that varies from 3 to 9 mm/yr along its length and is less than the estimated geologic slip rate of \(\sim 9\) mm/yr (Schmidt et al., 2005). The resulting slip deficit is expected to contribute to the occurrence of future large events (Lienkaemper et al., 2001; Schmidt et al., 2005).

Seismic strain accumulation is inferred from the difference between the far-field displacement rates and the interseismic creep rates along the fault surface. Thus, characterization of the creep rate distribution over time reveals variations in seismic strain accumulation rates that directly impact hazard characterizations. Previous studies of the relative motion along the Hayward fault have incorporated not only geodetic data from both continuous GPS and alinement arrays (Lienkaemper et al., 2001; dAlessio et al., 2005; Lienkaemper et al., 2012) but also processing techniques for differential InSAR (DInSAR) (Bürgmann et al., 2000; Bürgmann et al., 2006; Schmidt et al., 2005; Lanari et al., 2007). These results all suggest a complicated pattern of...
locking and creep at depth. Anywhere from 13 to 49 SAR images were employed to invert for deformation over time periods of eight or more years. DInSAR results have been limited by a lack of coherence along the hills on the eastern side of the fault.

In this paper we apply a new DInSAR technique (Samsonov and Tiampo, 2011) in order to produce a deformation map for a region along the northern Hayward fault using fifteen quad polarization RADARSAT-2 SAR images acquired during a three-year period from 26 April 2008 until 18 March 2011. This new method uses HH and VV polarized SAR images in order to select persistent scatterers (PS) based on their polarization phase difference (PPD). The normalized PPD index selects pixels that demonstrate predominantly even or odd bounce scattering properties, where odd bounce scattering is produced by reflection from a flat, rough surface and even bounce scattering is produced by the interaction of the radar wave with structures or other standing objects. The PPD approach requires a much smaller set of SAR acquisitions over a shorter period of time than other PS techniques. The result is a powerful technique to resolve potential PS candidates, which provide a more complete spatial distribution of pixels than alternative DInSAR techniques.

In previous work we compared the results for scatterers selected using the original PS technique (Ferretti et al., 2001) with those selected using the PPD technique and determined that the PPD technique successfully identified a larger percentage of pixels in areas of low coherence (Samsonov and Tiampo, 2011). Here we employ this new PPD technique to reliably estimate the deformation with fewer acquisitions and over a shorter time period than previous DInSAR studies, 2008–2011, with increased spatial coverage for better quantification of kinematic tectonic features and regional hazard estimates. These results suggest that shallow creep along the fault may be time-dependent on time periods of only a few years, supporting recent results from the alignment array (Lienkaemper et al., 2012). While longer term averages are most useful for seismic hazard assessments, those estimates primarily are based on geodetic measurements, including approximately 20 yr of GPS data and 15 yr of SAR images, supplemented by alignment data, tiltmeters and strainmeters, and campaign GPS, all with varying spatial and temporal coverage. If we designate these time periods as intermediate-term, and interseismic periods as long-term (on the order of 100 yr or more), the time periods examined here, ~3 yr, can be considered short-term. While slip rates based upon intermediate-term time periods are valuable input into seismic hazard estimates, variations at shorter time periods suggest that those estimates may be subject to larger variation than previously considered and has implications for fault dynamics, long- and short-term creep rates, and the associated seismic hazard estimates.

2. Data and methods

In a densely vegetated environment, two-pass DInSAR methodology is either limited or impossible due to severe temporal decorrelation. PS analysis has emerged as an effective alternative (Ferretti et al., 2001; Hooper et al., 2007; Hooper, 2008). The PS approach is based on the selection of pixels that are dominated by a predominant scattering mechanism that is consistent over a long period of time. Today, PS interferometry is an established technique that has proven successful for mapping ground deformation of various origins on a world-wide scale (Ferretti et al., 2001; Hilley et al., 2004; Hooper et al., 2007; Hooper, 2008). In PS analysis, persistent scatterers are selected based on their amplitude dispersion through time, where $D = A_0$, which is calculated as $D = \frac{A_0}{A}$, where $A$ is the standard deviation and $A_0$ is the mean amplitude calculated for the same pixel of the coregistered set of SAR images. However, in order to confidently select PS pixels with a high degree of accuracy it is necessary to have at least thirty SAR images (Ferretti et al., 2001). While large numbers of images provide stable deformation estimates with low noise levels, it is often difficult or expensive to acquire large numbers of images, and those images inevitably span large time periods, making it difficult to identify short-term variations.

Here we employ an improved polarization PS technique that does not require a large number of acquisitions for PS selection (Samsonov and Tiampo, 2011). This technique, applied to HH and VV polarized RADARSAT-2 SAR images, selects PS based on their PPD, $\Delta \phi$:

$$
\Delta \phi = \phi_{HH} - \phi_{VV}
$$

where $\phi_{HH}$ is the phase of a wave transmitted and received in horizontal polarization and $\phi_{VV}$ is the phase of a wave transmitted and received in vertical polarization, relative to Earth’s surface. Extreme values of PPD equal to zero and $\pm \pi$ correspond to scatterers with a predominant reflective mechanism, deterministic odd and even bounce scatterers. In most cases odd bounce scattering is produced by reflection from the flat, rough surface and even bounce scattering is produced by the interaction of the radar wave with anthropogenic structures or other standing objects (tree trunks, rocks, etc.). The PPD value diverges from
these extremes as the contribution from diffusive scattering increases (Evans et al., 1988; Ulaby et al., 1987). Diffusive scattering from vegetation produces PPD values that are randomly distributed in $[-\pi, \pi]$. In addition, we apply constraints on pixel amplitude to exclude single bounce scatterers from the water surface.

Here we use the normalized PPD index for the selection of reliable scatterers whose phase carries consistent information. This can be done by calculating for each pixel a normalized average of absolute values of PPD for a set of SAR images and by selecting pixels with values close to 0 and 1 (Samsonov and Tiampo, 2011):

$$\chi = \frac{\sum_{k=1}^{K} \Delta \phi_k}{K \pi}$$

where $K$ is the number of SAR images used for processing. In practice, threshold values of $\chi \leq 0.2$ and $\chi \geq 0.8$ are applied in order to select for scatterers dominated by odd and even bounce mechanisms, resulting in the selection of approximately 1.4% of the pixels for PPD DInSAR analysis (Samsonov and Tiampo, 2011). If we only investigate those pixels above a selected amplitude threshold that is chosen to remove pixels over the water surface and other areas of low-intensity backscatter, we recover approximately 131,000 pixels. From those ~82,000 pixels, or 62.5%, are the odd bounce scatterers and the remaining ~49,000 pixels, or 37.5%, are the even bounce scatterers. The $\chi$ threshold values 0.2 and 0.8 here are selected based on the trial and error approach and more advanced theoretical studies are underway.

Fig. 2 shows a subset of the original analysis over the waterfront and the surrounding hills. Pixels were selected by both a PS analysis (Fig. 2a) and a PPD analysis (Fig. 2b). In Fig. 2a the PS pixels are shown in blue while in Fig. 2b the PPD oddbounce pixels are shown in blue and the evenbounce pixels are shown in red. The PPD method selects a significantly greater number of pixels. In this case, the total number of PS pixels is approximately 107,000, while the total number of PPD pixels is approximately 132,000. Visual inspection confirms that these correspond to roads, bridges and open surfaces or rocks. Further details on pixel behavior with varying threshold values can be found in the Supplementary Material.

In this work, surface displacements near the Hayward fault south from Point Pinole to just north of the city of Hayward, CA (Fig. 1a) are calculated using the PPD technique from 15 descending HH and VV polarization RADARSAT-2 SAR (FQ7) images acquired over a 3 yr period between April of 2008 and March of 2011, detailed in Table 1. FQ7 is located as shown in Fig. 1b, with an approximate center latitude and longitude of 37.9° and −122.25°, respectively, a spatial resolution of approximately 5 m by 5 m, and an incidence angle of 28°.

The land cover in the imaged area consists of regions of dense vegetation, urban areas and open water. We selected 54 interferograms with baselines less than 400 m, where the bulk of the baselines are less than 250 m, and time spans ranging from 300 days to approximately 3 yr (see Supplementary Material, Section S2, for additional details on PPD estimation). Interferograms with shorter time spans were removed in order to increase the signal-to-noise ratio of the computed deformation rates. We initially performed standard DInSAR pre-processing for selected pairs using $5 \times 5$ multilooking that produced interferograms with limited coverage in vegetated regions but good signal-to-noise ratio in urban areas. These interferograms were used to constrain the orbital correction, to extract phases and topographic heights for a number of highly coherent pixels, and to perform a least squares adjustment of spatial baselines. Refined orbital parameters were used in the PPD processing.

The PPD processing employed data at the original resolution and removed the topographic phase using a 10 m National Elevation Database (NED) Digital Elevation Model (DEM) provided by the USGS. The phase was filtered using a spectral filter (Goldstein and Werner, 1998). Wrapped phases were extracted from the complex interferograms and used for time series analysis without phase unwrapping in order to correctly capture the discontinuity across the Hayward fault. Inspection of the

Table 1

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Fig. 2. (a) Pixels selected using a PS analysis over the waterfront and surrounding hills shown in blue and (b) pixels selected by a PPD analysis in the same region. Oddbounce pixels are colored blue, evenbounce pixels are colored red. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
interferograms showed that this approach is reasonable (see Supplementary Material). Phase jumps in spatial fringes caused by tropospheric noise were observed in areas remote from the fault region; however, they were clearly localized and random in space and time.

Subsequent time series analysis to estimate the linear deformation rates was performed using a method that is related to the Small Baseline Subset (SBAS) algorithm (Berardino et al., 2002; Schmidt and Bürgmann, 2003; Usai, 2003) and proposed by Samsonov et al. (2011). The time series are sparse because the analysis includes a limited number of images between 2008 and 2011. As a result, for this relatively small level of deformation we can only accurately estimate the linear deformation rate. The resulting least squares solution can be used to remove random atmospheric noise and estimate a linear trend, or velocity.

One of the most challenging problems in radar interferometry is presented by the correlated residuals that result from the modeling of satellite orbits. This is a particular concern when they occur in conjunction with multiple phase contributions in differential interferograms with similar spatial pattern and magnitude, as occurs here in conjunction with the tectonic interseismic strain accumulation on either side of the predominantly strike-slip Hayward fault. Recently, methods have been proposed to distinguish between those two components (Biggs et al., 2007; Gourmelen et al., 2010). While deformation features such as nodal lines and arctan profiles can be used to differentiate between various sources and magnitudes, both alternatives rely on either the estimation of a typically far field signal using long concatenated single SAR image frames (e.g., distance > 50 km normal to the fault) or the use of additional measurements (e.g., GPS time series results). In this analysis, additional images for concatenation were not available and our analysis is restricted to a region approximately 50 km square. In the present case, the RADARSAT-2 frame dimension reduces our ability to map the interseismic strain accumulation across the Hayward fault, which according to the inverse tangent model scales approximately with the elastic thickness of the crust (e.g., 9–12 km). Several different approaches were tried and are discussed in the Supplementary Material (Section S3). In the option presented here, we model the interseismic strain accumulation from all major faults in the region as detailed in Schmidt et al., 2005 and Section S3. After removal of the interseismic signal, the residual orbital trend was removed by estimating a 2-D linear model that explicitly accounts for the possible bias which can be introduced by the presence of a phase jump associated with the surface fault crept sections. We employed a bilinear model fit to the data on either side of the fault. We estimated model parameters for the northeast and southwest, and we averaged both. Then the average estimated parameters were applied to all the data points to ensure mutual agreement on both sides of the fault and that there was no bias introduced at the fault plane. The residual map then allows us to analyze the spatial variability of the shallow creep sections of the northern Hayward fault. We compared these results with independent geodetic data, short-wavelength alinement array estimates of creep on the fault trace (McFarland et al., 2009), and longer-wavelength GPS velocity estimates from local and regional arrays (d’Alessio et al., 2005), which initially were used to constrain the estimates of interseismic creep at depth.

A shaded relief map for the region in the dashed box of Fig. 1b, with the Hayward fault trace outlined and the theodolite alinement stations from McFarland et al. (2009), is shown in Fig. 3a. The resulting line-of-sight (LOS) PPD deformation rates between April 2008 and March 2011, analyzed to remove topography, atmospheric errors, deep interseismic creep and orbital ramp as described above, are shown in Fig. 3b. This image not only includes coverage of the region west of the fault that is comparable to that of other earlier work, it also provides estimates for the entire region to the east of the fault that previously was not possible using other methods for DInSAR processing (Schmidt et al., 2005; Bürgmann et al., 2006; Lanari et al., 2007). Displacements east of the fault are noisier than to the west, because lower coherence along the vegetated slopes increases phase noise. However, the fault trace is clearly outlined along the entire length and there appears to be a number of larger, spatially correlated regions. LOS deformation rates from the PPD DInSAR analysis (Fig. 3c) compare well with creep rates from alinement stations along the fault trace (McFarland et al., 2009), 2007–2010.

3. Modeling and analysis

To estimate the slip distribution on the fault plane, we discretized the fault into patches 2000 m in length by 1200 m in depth and applied a constrained least squares algorithm. The geometry of the fault was obtained from the alinement network locations, and the depth was fixed at 12 km (Bürgmann et al., 2000). Inversions using the PPD InSAR results alone produce high rates of creep on the shallow section of the fault (see Supplementary Material, Section S4), but produce significant residuals at longer spatial wavelengths. The final inversion process included adding back in the interseismic signal estimated from the same interseismic model discussed above and then inverting for deep creep on all regional faults and shallow creep along the Hayward, as detailed in the Supplementary Material (Section S4). Deformation was calculated based on a rectangular dislocation model (Okada, 1985), and the slip rate was constrained at less than 10 mm/yr. The inversion was tested for a series of smoothing parameter values, $\lambda$, which effectively dampen the wavelength of the slip spatial variation in the data. Here, $\lambda$ controls the degree of smoothing in the slip rate distribution. The optimal value was chosen as $\lambda=4$ (see Supplementary Material for details), and results are shown in Fig. 4.

The displacement field estimated from the combination of the preferred slip distribution model of Fig. 4d and the estimated interseismic slip, as detailed in the Supplementary Material (S4) is shown in Fig. 4b, while the residual between the data (Fig. 4a) and the model (Fig. 4b) is shown in Fig. 4c. Very little signal of any significance remains in the residual, although residual motion in the northeast may be related to the Concord fault. Small areas of LOS motion away from the satellite, along the fault scarp and further to the east, may be related to slow-moving landslides (Hilley et al., 2004).

Profile A–A’, Fig. 5a, compares the PPD DInSAR results from Fig. 4 with the continuous GPS estimates perpendicular to the fault plane (Fig. 1b). Blue dots are LOS deformation rates from Fig. 1b, while filled dots are interseismic GPS displacement rates within ± 25 km of the profile projected onto LOS. Modeled values from Fig. 4b are shown in red. The relative creep on the fault interface is distinct, at almost 2 mm/yr, and the fit is quite good within 15 km on either side of the fault. To the west of the Hayward fault, the model and LOS velocities diverge, likely due to unmodeled deep creep at depth on the San Andreas fault (0–3 km). At the northeastern end of the profile (37–40 km), there is a persistent residual that is not accounted for by the Hayward fault model or a linear interseismic velocity. Possible explanations include either activity along the Concord fault or local groundwater effects.

Seismicity within ± 5 km of the profile, $1 \leq M \leq 4.9$, is shown in Fig. 5b. Open dots are those events that occurred between 1990 and 2007, inclusive, while the filled dots occurred between 2008 and early 2011. Persistent seismicity occurs at the Hayward fault and the southwestern edge of the profile near the San Andreas fault. To the east, slightly deeper seismicity occurs along
the entire fault profile; however, seismicity that occurs during the 2008–2011 interval is clustered along the Hayward fault and at the Concord fault, where deformation residuals occur as noted above.

An independent comparison with the most recent revision of the surface creep rates from McFarland et al. (2009) is presented in Fig. 6. Auxiliary material reports linear rates for the period 1979–2010, a much longer period compared to our temporal DInSAR data, 2008–2011 (see Supplementary Material, Section S1). Creep rates (red squares) at the Hayward fault are calculated from the model of Fig. 4b and compared to creep rates for the time period 2008–2011 (black dots), with associated error bars. Here the model agrees quite well with the average alinement creep data, although higher values in the central section are likely a result of unmodeled fault complexity. Differences in the southern section may be due to additional short term variations in the measured creep which are masked by the three-year average, including a slow-slip event recorded in 2007 (Lienkamper et al., 2012). The results are not sensitive to the subsidence due to compaction of near-surface sediment deposits in the San Leandro basin (see Supplementary Material, S4).

We compare the corrected PPDDInSAR ground deformation from Fig. 3b and the model shown in Fig. 4b with the horizontal GPS velocities in Fig. 7. Fig. 7a shows the PPD data and the GPS vectors calculated by combining the long-term solutions of the BaVuGPS data (d’Alessio et al., 2005), and the PBO GPS velocity solution calculated through the end of 2010 (http://pbo.unavco.org/data/gps). We calculated the displacement rates at the same GPS locations from this model and compare them to the actual GPS velocities in Fig. 7b. The predicted values are a reasonably good fit to the estimated values. Exceptions include minor variations in rate and direction adjacent to the fault, which may...
Fig. 4. (a) Line-of-sight (LOS) linear deformation rates, 2008–2011, from a PPD analysis for the fifteen acquisitions listed in Table 1 and the addition of the deformation rate associated with regional interseismic creep at depth. Region is as shown in the dashed box, Fig. 1b. (b) Surface deformation rates estimated using the slip distribution model shown in (d), again added to the deformation rate calculated based upon regional interseismic fault creep at depth. (c) Residual velocity from (a) not accounted for in model (b). (d) Slip distribution model from inversion of deformation rates shown in (a), smoothing factor $k = 4$. Open dots represent seismicity between April 2008 and March 2011. The x-axis origin is aligned with Point Pinole, where black triangles denote alignment locations. (For interpretation of the references to color in this figure, the reader is referred to the web version of this article.)

Fig. 5. (a) Profile A–A, perpendicular to the fault plane, as in Fig. 1b. Blue dots are line-of-sight (LOS) deformation rates from Fig. 3b. Filled dots are continuous GPS displacement rates from PBO stations within $\pm 25$ km of the profile projected onto SAR LOS, with associated error bars. Red dots are modeled LOS velocities from Fig. 4b. (b) Seismicity within $\pm 5$ km of the profile, $1 \leq M \leq 4.9$. Open dots occurred between 1990 and 2007, inclusive; filled dots occurred between 2008 and early 2011. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
be associated with short-term variations in fault slip recorded by the DInSAR over this period. These are primarily associated with GPS campaign sites that display consistently higher errors than the bulk of the network (Schmidt et al., 2005). Differences further from the fault are most likely caused by unmodeled tectonic motion or deep slip on faults such as the Concord and Calaveras faults to the east, or deep slip along the San Andreas to the west that also is not included in this analysis.

4. Discussion

While the slip distribution profile of Fig. 4d does confirm earlier work suggesting that a complicated network of locked and creeping patches exists along the Hayward fault over the 2008 through 2011 interval, there are important features in the distribution that may help to resolve discrepancies between them (Bürgmann et al., 2000; Schmidt et al., 2005). For example, while this work also provides evidence for a large locked patch at depth at approximately 30 km along strike from Point Pinole (Fig. 4d), motion at depth occurs along most of the northern 20 km of the section, in accordance with the earlier work of Bürgmann et al. (2000), but decreases significantly near the surface. Detailed sensitivity analysis (Supplementary Material, S4) shows that this patch can be related to deformation located approximately 10 km east of the fault, on the northwest edge of the image. Due to the limitations associated with resolving slip at depth using surface data, it is not possible to determine from this work whether that is a direct response to creep on the Hayward fault at depths of 9–12 km below Point Pinole or if the locking depth near Point Pinole is shallower than further south. Another possible explanation is that motion on the Concord fault or local ground subsidence is mapped into creep at depth on the Hayward fault in the inversion.

These results are in accordance with the broader pattern of rates estimated from continuous GPS. This may be due to the identification of a substantially larger number of radar targets to the east of the fault at significantly larger distances, providing better resolution at multiple scales. Small patches of higher change rates adjacent to the fault likely correspond to local landslides, as seen in Hilley et al. (2004), and may be responsible for similar patches in the hills to the east of the fault. However, the alternating, broader rate change patches to the east, adjacent to the Calaveras fault, may be the result of tectonic motion on these secondary faults. While they are neither detailed nor significant enough to be modeled conclusively with this data alone, we did perform a number of inversions to study both their potential effect on the slip model of Fig. 3b, and to quantify the importance of the additional information provided by the increase in spatial coverage to the east provided by this PPD analysis (Figs. S29–S42, Supplementary Material). These results confirm that the extension of available data on the eastern side of the fault provide additional information on the slip pattern at depth. In particular, it is deformation between ten and fifteen kilometers east of Point Pinole provided by that additional coverage and which is not available to the west of the fault that is the source of the slipping region at depth at the northern edge of the fault plane. Variation between this result and that of others may be due in part to the different data sets and modeling approach. However, examination confirms that each side of the relatively symmetric deformation seen in the original image (Fig. 4a) contributes in part to the

![Fig. 6. Creep rates at the Hayward fault for the model of Fig. 4b (red dots) compared to creep rates (McFarland et al., 2009) calculated for the approximate time period 2008–2011 (black dots; see Section S1, Supplementary Material), with associated error bars. The x-axis origin is aligned with Point Pinole. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)](image)

![Fig. 7. (a) PPD DInSAR ground deformation from Fig. 3b and GPS vectors (black arrows) that combine the long-term solutions of BaVu network (d’Alessio et al., 2005) and the PBO GPS velocity solution through the end of 2010 (http://pbo.unavco.org/data/gps). (b) Predicted DInSAR ground deformation using the model solution shown in Fig. 4b and the predicted displacements rates from that same model at the GPS locations (red-tipped arrows). Grey arrows are again the long-term GPS rates as in (a). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)](image)
locking patches at 3–6 km depth and 20 and 30 km from Point Pinole in the full inversion.

These results for shallow slip along the northern Hayward fault raise the possibility that this short-term slip profile may be a snapshot of a longer and more complicated temporal pattern of motion. For example, recent work of de Michele et al. (2011), which analyzed 11 yr of data, concluded that spatially varying time-dependent motion occurs along the Parkfield section of the San Andreas fault further to the south, over periods as short as 1 yr and lengths of ~10 km. Analysis of more than 20 yr of alignment array data along the Hayward fault confirms that temporal variations in slip rate along the Hayward fault (Lienkaemper et al., 2012). The average slip rate of ~4.4 mm/yr across the profile for the 2008–2011 time period is lower than the corresponding rate for earlier estimates of longer time periods (Schmidt et al., 2005; Bürgmann et al., 2000, 2006; Malservisi et al., 2003); however, this study only encompasses the northern half of the fault. Additional studies of the entire fault length over a variety of time periods are necessary to constrain the magnitude and variation of the creep variation in time, and the associated temporal scales of that variation are necessary to constrain the spatial pattern of the long-term creep rate, the total strain accumulation, and the associated seismic hazard.

5. Conclusions

We present new results for the shallow fault motion along the Hayward fault using a new DInSAR technique, PPD analysis of RADARSAT-2 HH and VV polarization SAR images. This method provides, for the first time, the ability to map the displacement field on both sides of the fault, including the vegetated hills to the east. Although the additional information contained in the data, this analysis can be performed with only fifteen polarimetric acquisitions over only 3 yr, less than half as many as previous DInSAR studies. This results in a much shorter time period than available in the past at approximately half the cost. We use the results to estimate short-term shallow creep and with better spatial coverage on both sides of the fault, and validate it against long-wavelength GPS data and short-wavelength surface creep data along the fault itself. Although the motion at depth is better constrained by the additional PPD pixels identified to the east of the fault (see Supplementary Material), the associated shallow fault slip model not only provides a good fit to the regional continuous GPS velocities, but also to the long-term creep rate at the surface.

Inverse modeling of the slip distribution along the fault confirms that there is a large locked patch at depth, 30 km along strike, but it also identifies with better resolution creeping areas at different depths and further north along strike. The fault is slipping at depths below 6 km for the northernmost twenty km of the fault, similar to the results of Bürgmann et al. (2000); however, in the same area there is a persistent locked region above 6 km evident in most of the inversions. These new details and the differences from earlier solutions are likely the result of two significant improvements intrinsic to this new PPD DInSAR technique. First, the time period evaluated here, slightly less than 3 yr, is significantly shorter than that of earlier DInSAR studies along the Hayward fault. Application of PPD analysis to dynamic phenomena helps to resolve shorter temporal patterns of deformation caused by the time-dependent variations in slip along the fault. Here we estimate a linear velocity over a short time period. Accurate estimates of the variation over longer time periods will require analysis of additional images over varying time periods. Second, this technique allows us to image large areas on the eastern side of the fault that could not be imaged previously due to lack of coherence in the region. Sensitivity studies suggest that this additional coverage provides information critical to our understanding of shallow aseismic fault slip at all depths, although more detailed studies are necessary to definitively identify the appropriate sources.

Although the current availability of quad polarization data is significantly less than that of other acquisition modes of SAR data, and RADARSAT-2 images generally are expensive, the success of the PPD algorithm provides a new and innovative advance in the field of DInSAR.

While additional work is necessary to determine whether the revised slip profile seen here is due to the better resolution provided by the PPD analysis, the expected variation arising from the combination of different types of geodetic data, or short term temporal variations in that slip rate, the implications to seismic hazard estimation are significant, suggesting that the earthquake potential in the area may be less than anticipated from earlier work. Although we only image the northern Hayward fault in this work, Lienkamper et al., (2012) have documented temporal variations on the order of 10–25% along the central and southern Hayward that suggest that the magnitude and frequency of large events may depend on processes within the fault zone itself. Evidence that the interseismic rate varies over short time periods prompts us to ask whether we are witnessing a temporary decrease in the creep rate or if that decrease is part of a longer term spatial variation in slip along the entire fault. The correct answer has important implications for our understanding of fault dynamics, long- and short-term interseismic creep rates, and the associated hazard estimates. Identifying the relative significance and contribution of each of the above improvements to the final solution will require analysis of additional images, including an evaluation the magnitude and scope of the spatial and temporal variations in creep and locking depth along the fault plane, the sensitivity of that distribution to the spatial pattern of the DInSAR displacement field, and incorporation of other sources into a more realistic model. However, this work has demonstrated the potential impact of this method in defining the spatial and temporal complexity of aseismic slip and accumulated elastic strain for one of the most significant sources of seismic hazard in the San Francisco Bay area.

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Appendix A. Supporting information

Supplementary data associated with this article can be found in the online version at http://dx.doi.org/10.1016/j.epsl.2013.02.019.

References

