

1 **Geodetic Evidence for Distributed Flow Below the**
2 **Brittle Crust of the Walker Lane, Western United**
3 **States**

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7 **Key Points:**

- 8 • Geodetic velocities in the Walker Lane (WL) reflect distributed shear in the lower
9 crust rather than deformation due to discrete faults.
10 • The width of and velocity across the northern WL is 172 ± 6 km and 7.2 ± 0.3 mm/yr,
11 resp., and 116 ± 4 km and 10.1 ± 0.2 mm/yr for the central WL.
12 • Estimating fault slip rates using models that assume their downdip continuation
13 into the lower crust may be inappropriate for some regions.

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Abstract

The predominant approach for modeling faults in the Earth’s crust represents them as elastic dislocations, extending downdip into the lower crust, where the faults slip continuously. The resulting surface deformation features strain accumulation concentrated across locked faults during the interseismic period. An alternative model proposes faults confined to the elastic crust, with surface deformation driven by a wide zone of distributed shear underneath. Using high-precision GPS data, we analyze deformation profiles across the Walker Lane (WL), USA. The WL is a transtensional region of complex faulting, which delineates the western edge of the Basin and Range province and accommodates a significant portion of the Pacific-North American plate boundary deformation budget. Despite a dense geodetic network surveyed collectively for nearly 20 years, horizontal velocities reveal no evidence of localized strain rate accumulation across fault surface expressions. Instead, deformation within the shear zone is uniformly linear, suggesting that the surface velocities reflect distributed shear within the ductile crust rather than discrete fault deformation. This implies no downdip fault extension below the seismogenic layer. The shear zone, bound by the Sierra Nevada crest in the west, is 172 ± 6 km wide in the northernmost WL narrowing to 116 ± 4 km in the central WL. This study’s conclusion challenges the assumption of the presence of dislocations in the lower crust when estimating geodetic slip rates, suggesting that slip rates are instead controlled by the fault’s position and orientation within the shear zone. This has important implications for quantifying seismic hazards in regions with complex fault systems.

Plain Language Summary

Interpreting Earth’s surface deformation, measured by high-precision GPS stations, is crucial for understanding plate tectonics and assessing seismic hazard. Traditionally, the assumption has been that faults in the Earth’s upper crust extend as discrete dislocations into the lower crust. In this paper, we show that there is no compelling evidence of this in the Walker Lane region of California and Nevada. Instead, we conclude that the geodetically measured deformation on the surface reflects uniform shearing in the lower crust. Our findings support the interpretation of the Walker Lane region as a developing large-scale strike-slip fault and imply that the current method of estimating slip rates on the faults may be inappropriate.

1 Introduction

A long-standing concept in tectonic geodesy is that of an elastic dislocation model (EDM), in which a fault is represented as a locked dislocation in the upper crust and with a continuously creeping continuation into the viscoelastic lower crust. For vertical faults, the EDM predicts an arctangent shape of the horizontal surface velocity field (Savage & Burford, 1973), resulting in localized shear strain on the surface across the fault trace. These signals can be detected in investigations of active crustal deformation, accessible through geodetic techniques such as InSAR (e.g. Wright et al., 2001; Tong et al., 2013; Cakir et al., 2014; Chaussard et al., 2016; Weiss et al., 2020), GNSS networks (e.g. Wdowinski et al., 2004; Meade & Hager, 2005; Schmalzle et al., 2006; Vernant, 2015; Hussain et al., 2018), alignment and leveling arrays (e.g. Savage et al., 1979; Galehouse & Lienkaemper, 2003; Mongovin & Philiposian, 2021).

A viscoelastic dislocation model (VEDM) assumes the same structure as the EDM, but takes the coupling between the viscous and the elastic layers into account (Savage & Prescott, 1978; Savage & Lisowski, 1998; Savage, 2000; Pollitz et al., 2008). The implication of VEDM is time-dependent strain rates, with a flattening of the arctangent shape late in the fault’s seismic cycle.

62 A competing concept is that of the shear zone model (SZM) (Prescott & Nur, 1981;
 63 Bourne et al., 1998; Pollitz, 2001), in which faults only exist in the elastic part of the
 64 crust, with the ductile layer underneath deforming smoothly without discrete disloca-
 65 tions. In this model, the surface velocity pattern is mostly linear (i.e., constant shear strain
 66 rate), reflecting the underlying shear. The elastic layer acts as a smoothing filter, broad-
 67 ening the expression of the shear zone and making the surface deformation more distributed
 68 with increasing thickness of the elastic layer.

69 The EDM has gained popularity, in part, due to its simplicity and utility in the es-
 70 timation of slip rates on the faults (e.g. Fay & Humphreys, 2005; Schmalzle et al., 2006;
 71 Hill & Blewitt, 2006). The deformation across large-scale strike-slip faults generally ex-
 72 hibits the arctangent shape and is fit well by the EDM or by the VEDM (Chuang & John-
 73 son, 2011; Vernant, 2015; Y. Zhu et al., 2020). Studies of exhumed peridotite massifs,
 74 ophiolites, and xenoliths (e.g. Norris & Cooper, 2003; Titus et al., 2007; Vauchez et al.,
 75 2012, and references therein), and seismic imaging and anisotropy (L. Zhu, 2000; Vauchez
 76 et al., 2012; Ford et al., 2014) further support the continuation of large strike-slip faults
 77 into the ductile portion of the lithosphere. However, each of the study methods has short-
 78 comings (Vauchez et al., 2012), preventing an unequivocal conclusion. Studies on the same
 79 fault zone can yield conflicting results. For instance, Titus et al. (2007) find that observed
 80 seismic shear wave splitting in central California is consistent with a broad shear zone
 81 in the upper mantle beneath the San Andreas transform, while seismic imaging done by
 82 Ford et al. (2014) supports a zone of localized shear (less than 50 km).

83 Another example of a major transform is the Alpine fault in the South Island of
 84 New Zealand. Despite evidence from exhumed xenoliths and massifs for localized shear
 85 underneath the Alpine fault (e.g. Norris & Cooper, 2003; Kidder et al., 2021), geophys-
 86 ical evidence is less conclusive. Moore et al. (2002) use seismic shear wave splitting to
 87 conclude that the wide shear deformation on the surface mirrors that at depth. Lamb
 88 and Smith (2013) find that the surface velocity in South Island is fully explained by the
 89 deep slip on the main Australian and Pacific plate interface; the data does not require
 90 deep creep beneath individual faults in the region. For both the San Andreas fault zone
 91 in Southern California and the Marlborough fault zone in the South Island of New Zealand,
 92 an extension of the Alpine fault, Bourne et al. (1998) suggest that the surface velocity
 93 can be accounted for by a distributed shear zone below, without faults extending into
 94 the ductile region of the lithosphere. Consequently, the questions of whether faults ex-
 95 tend beneath the brittle crust and whether the shear zone is localized or distributed re-
 96 main unresolved.

97 The vast majority of studies addressing these questions have focused on large-scale
 98 continental transforms, with few investigations on smaller faults situated away from the
 99 immediate vicinity of continental boundaries. In this paper, we evaluate geodetic defor-
 100 mation across the Walker Lane, in California and Nevada, USA, a region that is not a
 101 major continental transform. We apply a quantitative analysis to compare the two com-
 102 peting models, addressing data uncertainties to identify significant parameters. We find
 103 little support for the presence of dislocations in the viscoelastic layer of the lithosphere
 104 and present strong evidence in favor of distributed shear deformation beneath the elas-
 105 tic layer, with faults terminating within the brittle crust.

106 2 Tectonic Setting

107 The Walker Lane (WL) (Figure 1a) is an elongated zone of both shear and exten-
 108 sional deformation in eastern California and western Nevada, separating the Sierra Nevada
 109 mountain range to the west from the Basin and Range Province to the east. The WL
 110 accommodates a substantial part of the relative active motion between the Pacific and
 111 North American plates (Bennett et al., 2003; Hammond et al., 2011). It is a dynamic
 112 and geologically complex region, exhibiting diverse topography and a variety of fault-

ing styles in complex network. It has been speculated that the WL is an immature continental boundary and will possibly become the main transform boundary in the future (Faulds et al., 2005; Wesnousky, 2005a; Pierce et al., 2021). Its northern section is characterized by northwest-striking, roughly parallel right-lateral strike-slip fault systems and northeast-trending left-lateral strike-slip faults (Wesnousky, 2005a). The deformation is predominantly shear in the region (Svarc et al., 2002; Hammond & Thatcher, 2004; Kreemer et al., 2009; Wesnousky et al., 2012; Kreemer & Young, 2022), associated with the translation of the Sierra Nevada/Central Valley microplate to the northwest with respect to the Basin and Range (Dixon et al., 2000; Argus & Gordon, 2001). The central WL, spanning from Walker Lake basin to Lake Tahoe basin, is characterized by a conspicuous absence of strike-slip faults (Wesnousky et al., 2012), with the exception of small north and northwest-trending strike-slip systems on the eastern side of the WL (Wesnousky, 2005a; Surpless, 2008; Dong et al., 2014; S. J. Angster et al., 2019; Pierce et al., 2021). A significant part of the motion in the central WL is accommodated by rotating crustal blocks and basins bounded by normal faults (Wesnousky et al., 2012; Bormann et al., 2016; Pierce et al., 2021).

3 Data Analysis

Modern high-precision GPS data achieve remarkable position accuracy (Blewitt, 2015; Bock & Melgar, 2016), which we improve to sub-millimeter levels through applying rigorous station selection criteria and position time series filtering. We use position time series in a North American plate reference frame, obtained from Nevada Geodetic Laboratory (Blewitt et al., 2018) and derived using the Precise Point Positioning method (more details in Kreemer et al. (2020)), using the GipsyX software by the Jet Propulsion Laboratory (JPL), and using JPL’s final GPS orbits and clocks (Bertiger et al., 2020). The majority of the data were collected through the MAGNET GPS network, which utilizes a semi-permanent methodology (Blewitt et al., 2009), supplemented by data from continuously operating stations, mostly from the EarthScope Network of the Americas, but also from the Washoe County GPS Network and Leica SmartNet Network (Figure 1a).

We consider all time-series in the period 2007.0-2023.0 that span at least 2.5 years. We apply a station motion model to the time-series that includes annual and semi-annual sinusoidal signals, accounts for offsets, and iteratively removes outliers defined by $> 3\sigma$ deviation in the residual time-series. Offsets are obtained from the list of potential discontinuities from GNSS equipment changes earthquakes available at the Nevada Geodetic Laboratory (<http://geodesy.unr.edu/NGLStationPages/steps.txt>). Accidentally unrecorded or erroneously introduced offsets in position-time series can result in larger errors in velocities, especially for the semi-continuous stations. We meticulously screen each station for unrecorded equipment offsets and assess the impact of nearby earthquakes on the data. Earthquake-related offsets are introduced only when there is clear evidence that the station had been affected in a manner consistent with the earthquake’s mechanism.

The time-series may sometimes be affected by non-tectonic processes, specifically hydrologic loading, which is more substantial for stations in the Sierra Nevada compared to those in the Great Basin. Not accounting for those signals can have an adverse effect on the station velocity, particularly for the semi-continuous stations. To best remove those signals, we apply a local common-mode filter to the data, using the method of Kreemer and Blewitt (2021), which effectively removes non-secular signals, leading to improved velocity estimates and smaller velocity uncertainties. In this method, only stations with >2000 position estimates are considered as filter stations (i.e., essentially the continuous stations and some of the frequently observed MAGNET stations) unless their residual time-series are not representative of the regional common-mode (see Kreemer and Blewitt (2021) for details) (Figures 1b and 2). Finally, we use MIDAS, a robust median

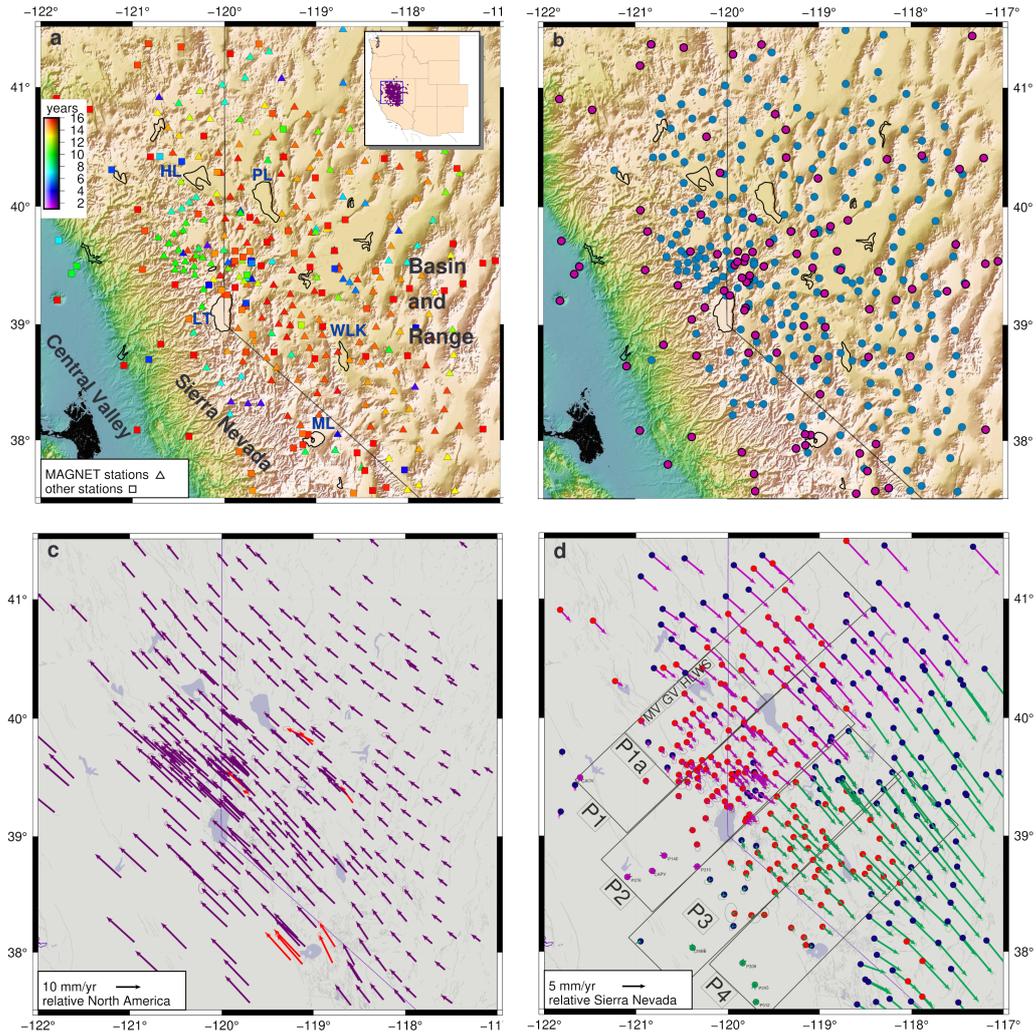


Figure 1. (A) Topographic map of the northern and central Walker Lane, showing major geologic features and lakes (Honey Lake – HL, Pyramid Lake – PL, Lake Tahoe – LT, Walker Lake – WLK), and the GPS stations (MAGNET stations are triangles), color-coded by the length of the time series. The inset shows the location of the map in the western United States and the stations used in the data processing and analysis (purple dots). (B) Stations used for filtering the GPS time series (purple dots) versus other stations (blue). (C) Shows the velocity field in a North America reference frame and the GPS stations omitted from the analysis (red vectors). (D) Map showing the two sets of reference stations (green and purple dots), the velocity field in the Sierra Nevada reference frame (color-coded with the reference stations), the strike-slip faults in the northern Walker Lane (MV – Mohawk Valley, GV – Grizzly Valley, HL – Honey Lake, and WS – Warm Springs), and the four profiles P1(a), P2, P3, and P4. For each profile, zero is defined as the western edge of the profile. The station dot color denotes the bounds of the shear zone based on the deviation of station velocity azimuth from the rotation field of the Sierra Nevada (red – less than 1.5° difference).

165 trend estimator (Blewitt et al., 2016), on the filtered and offset-corrected time-series to
 166 obtain each station’s velocity and its uncertainty. The velocity field relative to North Amer-
 167 ica is shown in Figure 1c for the 368 stations considered. Some stations with outlier ve-
 168 locity (typically observed for stations near active geothermal production areas) are iden-
 169 tified and excluded from the remaining analysis. Data Set S1 contains the velocities used
 170 henceforth.

171 The velocities used in this analysis are not corrected for postseismic relaxation. Post-
 172 seismic response of the viscoelastic lower crust and upper mantle following large earth-
 173 quakes can last tens to hundreds of years and can affect geodetic velocities (Nur & Mavko,
 174 1974; Savage & Prescott, 1978; Hammond et al., 2009). We explored the impact of cor-
 175 recting the velocities on our results in Supplemental Materials. While there are some dif-
 176 ferences, using the corrected velocity field yields the same conclusions that we are pre-
 177 senting here.

178 4 Modeling

179 4.1 Geodetic Profiles Across The Shear Zone

180 The Sierra Nevada (SN) west of the WL has been previously shown to have little
 181 internal deformation (Argus & Gordon, 1991; Dixon et al., 2000; Bennett et al., 2003;
 182 McCaffrey, 2005; Kreemer et al., 2009) and, therefore, provides a natural reference frame
 183 in which to analyze the velocity field across the WL. For this purpose, we use several long-
 184 term continuously-operating stations located on the rigid SN block to rotate the veloc-
 185 ity field into a SN reference frame (Figure 1d). In doing so, we find that the residual mo-
 186 tion of the SN sites is best reduced if we consider two distinct sets of reference stations,
 187 a northern and central set of five and four stations. By breaking up the SN into two dif-
 188 ferent reference blocks, we also insure that our profiles across the WL optimally cover
 189 the area. We subsequently estimate two Euler poles, one for each group of stations, and
 190 use them to create two different reference frames. That is, we use those poles to rotate
 191 the original velocity field in the northern part of our study area into a northern SN fixed
 192 reference frame, and the central part into a central SN frame (Figure 1d). If the veloc-
 193 ity field reflects shear in the WL, one would expect the SN fixed velocities to be paral-
 194 lel to small circles around the Euler poles. Because we expect stations in the east to start
 195 to reflect Basin and Range extension, we use the deviation of station velocity azimuths
 196 from the small circle azimuth within a specified tolerance as an estimate for the east-
 197 ern boundary of the shear zone.

198 To examine the deformation, we define four profiles across the northern and cen-
 199 tral WL, labeled P1 in the north through P4 in the south (Figure 1d), oriented such that
 200 the along-profile components of velocity within the shear zone, defined by the velocity
 201 azimuths (Figure 1d, red dot stations), are minimized. We use the northern SN frame
 202 for P1 and P2 and the central SN frame for P3 and P4. The velocity profiles are then
 203 obtained by projecting the velocities onto the profile-normal orientation.

204 The signals across all four of the resulting profiles are similar in shape, with uni-
 205 formly increasing across-profile velocities (Figure 2, blue dots) tapering off on the SN
 206 in the west and the Basin and Range in the east. There are no obvious inflections which
 207 might imply locations of the faults. The along-profile velocity component (Figure 2, black
 208 dots) is essentially zero in P1 and P2, but has a small signal in the eastern sections of
 209 P3 and P4, likely due to the narrowing of the shear zone in the southward direction.

210 4.2 Elastic Dislocation Model

211 According to the EDM for vertical strike-slip faults, the profile velocity, v , is a func-
 212 tion of the distance along the profile, x , the location of fault i in the profile, f_i , the slip

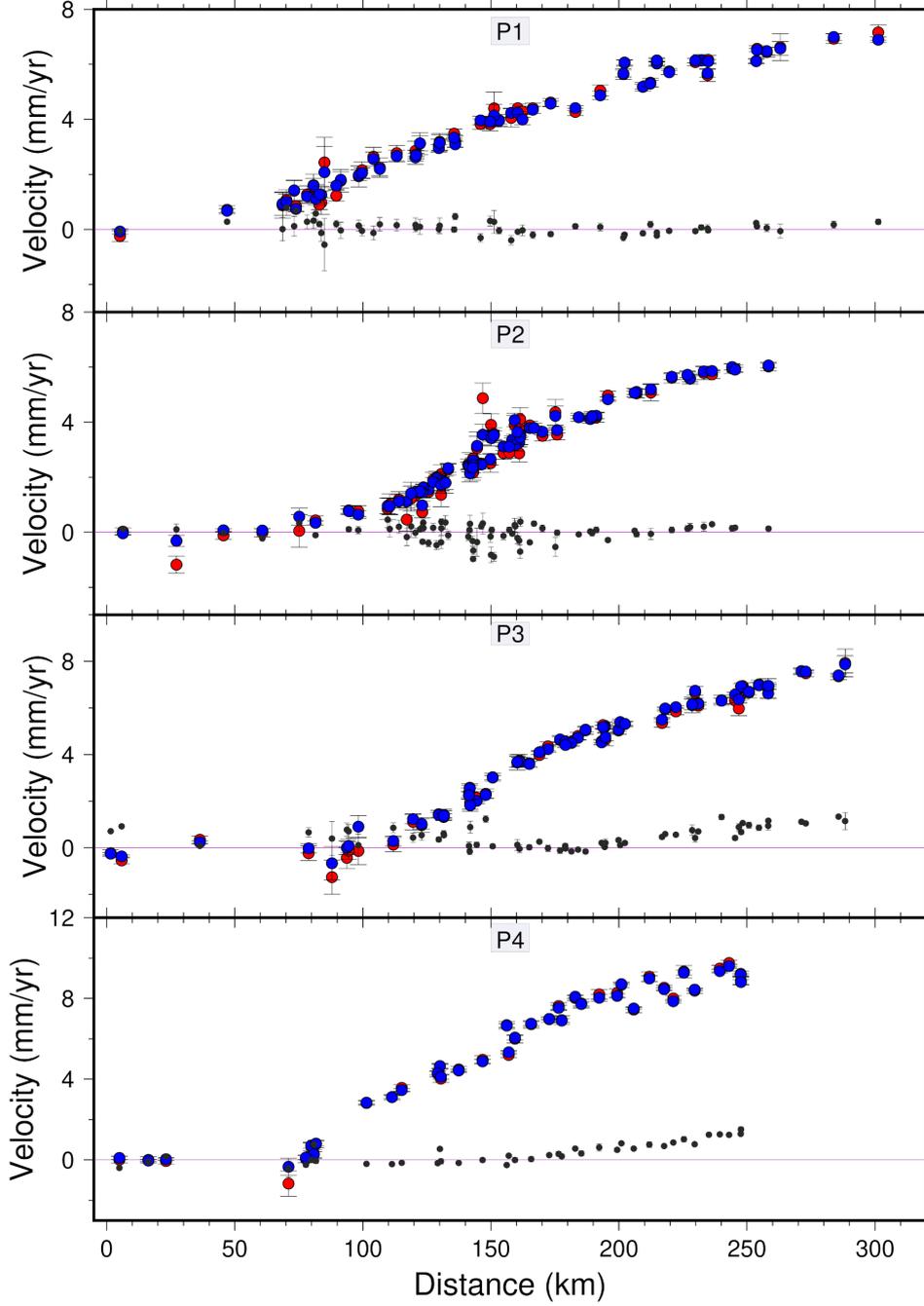


Figure 2. Velocity profiles P1, P2, P3, and P4 before (red) and after (blue) filtering. The outlines of the profiles are shown in Figure 1d. Note that the sign of the velocity is positive to the southeast in the SN frame. The velocity components that are along the direction of profiles are also shown relative to the northeast direction (black). The along-profile velocities are statistically not different from zero in P1 and P2 ($p > 0.05$), but have a trend in P3 and P4 ($p < 0.05$). This is due to the profiles being oriented to minimize the velocity vectors within the shear zone.

213 rate, s_i , and the locking depth, D_i , of each fault (Savage & Burford, 1973). Thus, we eval-
 214 uate the EDM using the equation

$$215 \quad v = \sum_{i=1}^N \frac{s_i}{\pi} \arctan\left(\frac{x - f_i}{D_i}\right) \quad (1)$$

216 where N is the number of faults present in the profile.

217 Since the four strike-slip faults we are interested in are all on the western side of
 218 P1 (Figure 1d), we define a subset profile P1a, beginning and terminating to the west
 219 and east, respectively, of the set of the strike-slip faults (Figure 3). There are other, pre-
 220 dominantly normal, faults in the eastern half of P1. Deformation due to dip-slip dislo-
 221 cations would have gradients in both strike-parallel and perpendicular velocity compo-
 222 nents. We do not observe a gradient in velocity when moving across the normal faults
 223 (Figure 2, black dots). Therefore, there is no need to consider any dislocation that would
 224 produce a strike-normal gradient, such as the north-south trending normal faults located
 225 in the eastern half of P1.

226 We combine the HL and WS faults into a single dislocation, HLWS, since the two
 227 are separated by only about 4-10 km. When the width of a deformation zone is less than
 228 the critical length, πD , the surface velocity due to any number of dislocations within the
 229 zone appears equivalent to that of a single dislocation, the one accommodating the ma-
 230 jority of the total slip (Traoré et al., 2014). Consequently, it is impossible to differen-
 231 tiate between distributed shear and a single dislocation over such areas (Moore et al.,
 232 2002). In our case, the critical length is ~ 50 km. Profile P1a extends over a length of 115
 233 km, theoretically allowing us to resolve at least two faults. We consider the MV fault
 234 on the western side of the profile and the combined HLWS faults on the eastern side as
 235 the two dislocations. This is similar to, e.g., the geodetic block model of Hammond et
 236 al. (2011).

237 We constrain each fault’s locking depth, D_i , using the seismogenic depths (Ruhl
 238 et al., 2020; Zuza & Cao, 2020): 17km for MV and 14 km for HLWS faults. We consider
 239 the surface trace of the faults along with relocated seismicity clusters (Figure 3) in con-
 240 straining the fault locations, f_i . We position MV fault above the obvious seismicity clus-
 241 ter and HLWS – approximately between the surface traces of the HL and WS faults. Af-
 242 ter constraining D_i and f_i , Equation 2 for the EDM is linear in the remaining param-
 243 eters, the slip rates s_i . Thus, we use a weighted linear least-squares approach to approx-
 244 imate the slip rates for the two dislocations.

245 4.3 Shear Zone Model

246 We adapt the parameterization of the SZM as described by Prescott and Nur (1981)
 247 and Prescott et al. (1981) (Figure 4). The strain field present on the surface above the
 248 shear zone is approximated by a distribution of infinitesimal screw dislocations. By in-
 249 tegrating the strain field over the width of the shear zone, the surface velocity is obtained
 250 as a function of the distance across the fault x , the velocity difference across the shear
 251 zone b (i.e., total slip rate), the thickness of the elastic layer D , and the half-width w
 252 of the shear zone below depth D (Prescott et al., 1981):

$$253 \quad v = -\frac{b}{2\pi w} \left[(x' - w) \arctan\left(\frac{x' - w}{D}\right) - (x' + w) \arctan\left(\frac{x' + w}{D}\right) \right. \\ 254 \quad \left. - \frac{D}{2} \ln\left(\frac{D^2 + (x' - w)^2}{D^2 + (x' + w)^2}\right) \right] + bc \quad (2)$$

255 where $x' = x - aw$. To determine the best-fit values for the parameters, we fit Equa-
 256 tion 2 to the four profiles—P1, P2, P3, and P4—employing a weighted nonlinear least-

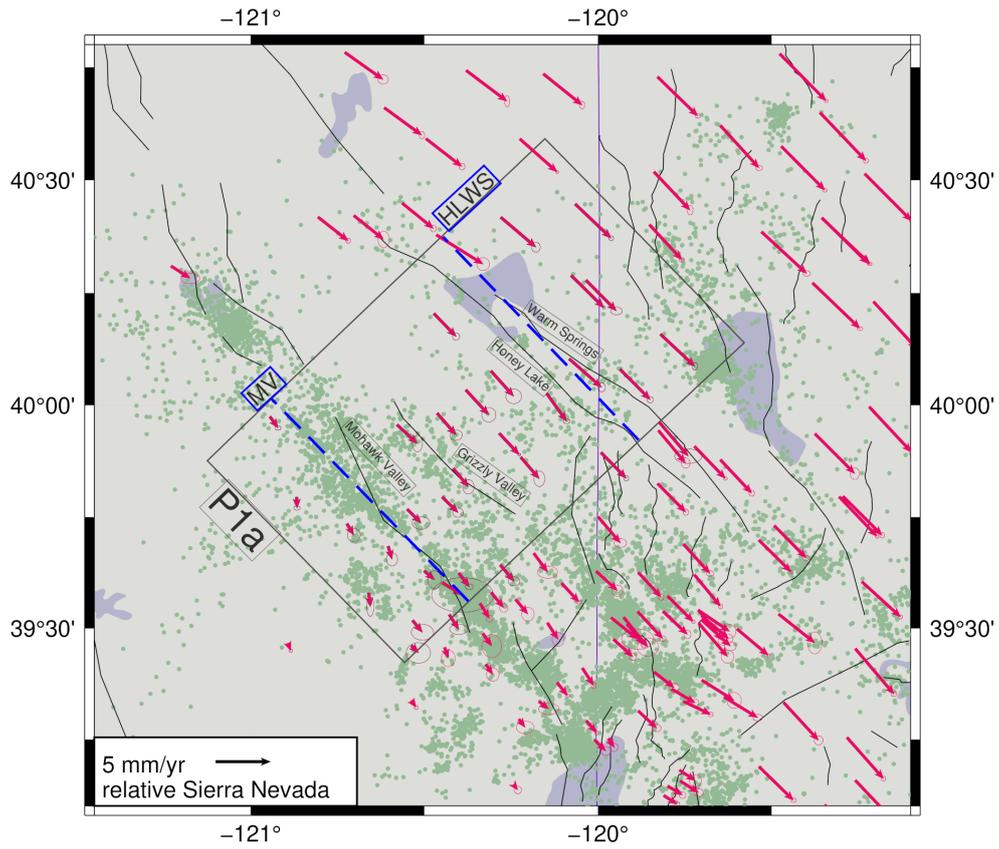


Figure 3. Map of the Northern Walker Lane, enlarged on profile P1a (gray box). Assumed Mohawk Valley (MV) and Honey Lake-Warm Spring (HLWS) fault locations are shown in blue dashed lines. The actual traces of the four strike-slip faults in the profile are also marked. Relocated seismicity (Ruhl et al., 2020) with $m \geq -1$ between 2002-2019 is shown in light green.

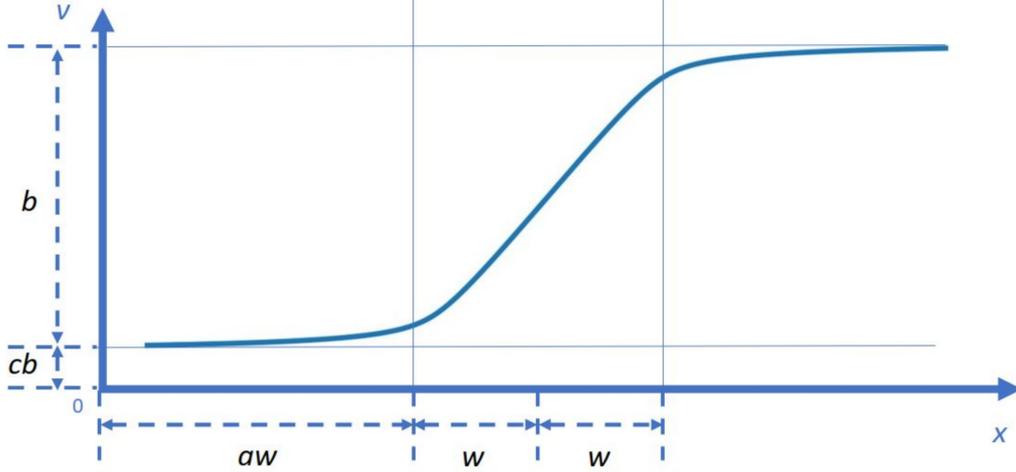


Figure 4. A diagram illustrating how Equation 2 characterizes surface deformation. The shear zone’s half-width (w) is located at depth D , and b is the total velocity difference across the shear zone. Dimensionless factors a and c align Equation 2 with the location of the western edge of the shear zone within the profile (a shifts the function horizontally and c shifts it vertically).

257 squares approach. We tested values of for w and D in a grid search and plot RMS mis-
 258 fit values for each profile in Figure 5.

259 A trade-off exists between the parameters D and w due to the coupling between
 260 the elastic and the viscoelastic layers. A combination of a large value of D and a small
 261 value of w can result in surface velocities similar to those due to a small value of D and
 262 a large value of w . To constrain this problem, either D or w must be determined through
 263 alternative sources of data. The RMS contour plots reveal that the half-width param-
 264 eter is well constrained by the data, while the depth is not. Relocated seismicity in the
 265 northern WL (Ruhl et al., 2020) indicates that the majority of seismic activity in the
 266 region takes place above 20 km. We therefore fix the upper bound for the value of D in
 267 Equation 2 to be 20 km, which is also near the minimum in RMS misfit at each profile
 268 (Figure 5). We use the shear zone width predicted by the velocity azimuths (Figure 1d)
 269 to compare with that predicted by the SZM, thus employing two independent methods
 270 for determining the shear zone bounds.

271 5 Results

272 The EDM fit to Profile P1a (Figure 6b) yields slip rates of 2.7 ± 0.1 mm/yr and
 273 2.2 ± 0.1 mm/yr for MV and HLWS faults respectively. The sum of the predicted slip
 274 rates is 4.9 mm/yr, which is a 68% of the 7.2 mm/yr relative velocity budget observed
 275 across the entire shear zone in P1. This discrepancy does not necessarily favor one model
 276 over the other, since profile P1a is a subset of P1 and has its own relative budget of about
 277 5 mm/yr across it. We do not fit the EDM to the entire shear zone for reasons described
 278 in section 4.2. The linear model-predicted strain rate across the zone (slope of the line)
 279 is 37 nanostrains/yr, which is in agreement with the shear strain calculated by Kreemer
 280 and Young (2022). We obtain excellent SZM fits for each of the four profiles (Figure 7)
 281 and find good agreement between the model-predicted shear zone width, w , and that pre-
 282 dicted by the velocity azimuths. The width of the shear zone is estimated to be the widest,
 283 172 ± 6 km, in the northern end of the WL. It then narrows to 130 ± 4 km near Lake

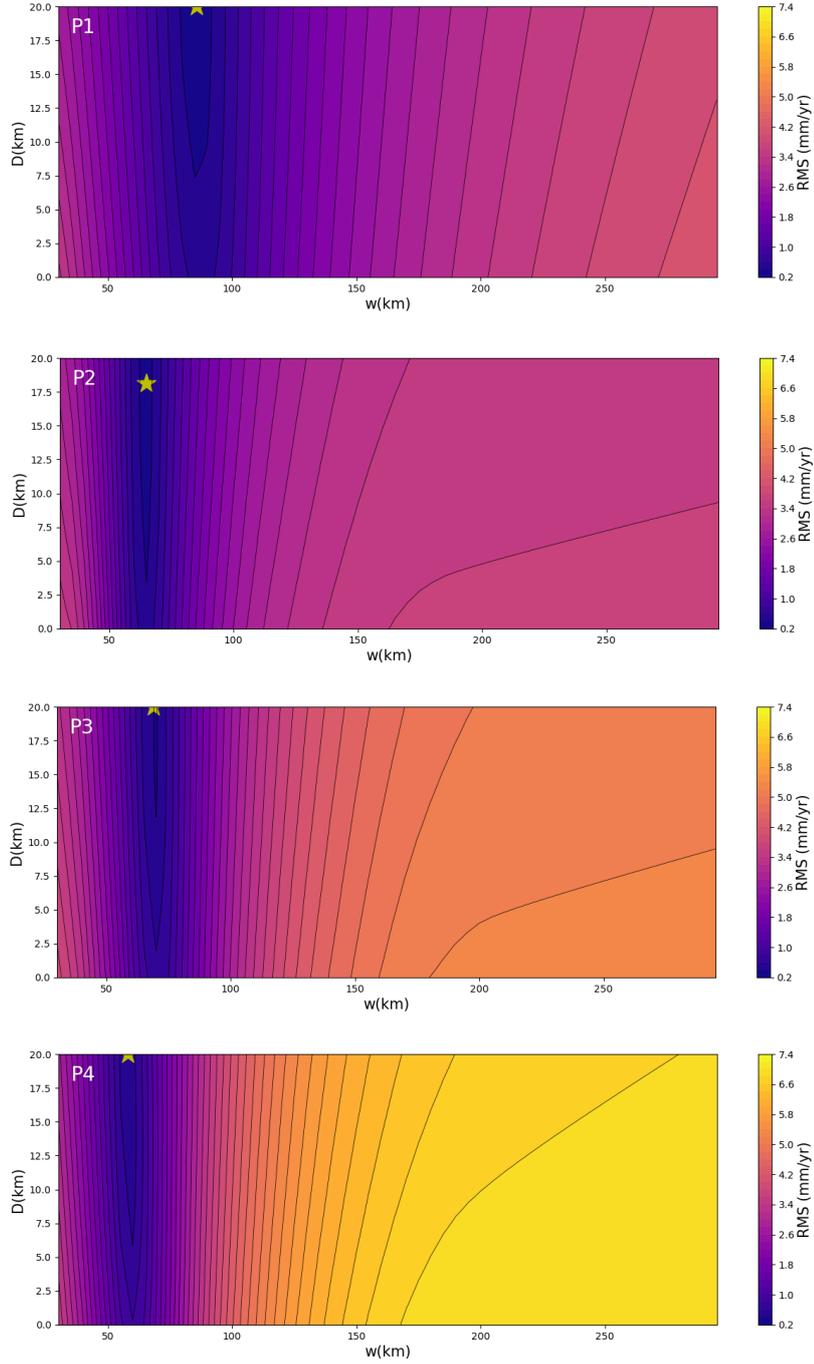


Figure 5. RMS of residual velocity data contour plots for the thickness of the elastic layer (D) and the half-width of the shear zone (w) in the Shear Zone Model (Equation 2) for profiles P1, P2, P3, and P4. The gold stars denote the best-fit results when we solve for both parameters, but having set an a priori maximum depth to 20km).

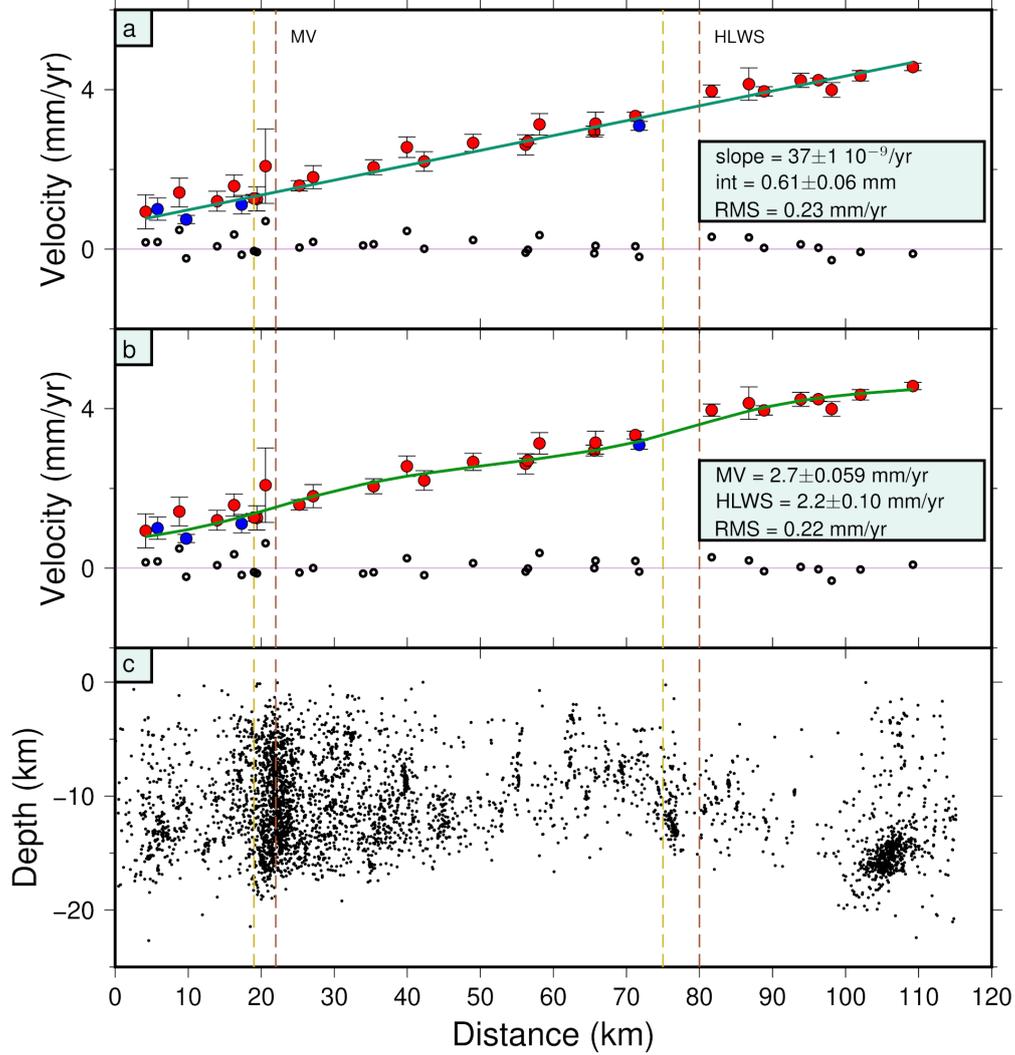


Figure 6. Profile P1a fault-parallel velocities are shown with the linear model (a) and the elastic dislocation model (b) fits. The locations of the faults are indicated by the dashed lines: brown – constrained using the surface fault traces, yellow – obtained by minimizing the misfit to the Elastic Dislocation Model (EDM) fit (Equation 1). The locking depths are constrained to be equivalent to seismogenic depth. The predicted parameters for the line and the EDM (slip rates of each fault) are listed. Relocated seismicity (Ruhl et al., 2020) with $m \geq -1$ between 2002-2019 is shown in the bottom panel (c).

284 Tahoe and 138 ± 6 km south of Lake Tahoe, before further narrowing to 116 ± 4 km
 285 (Figure 7, vertical dashed lines). Profiles P1, P3, and P4 are best-fit with a depth of 20
 286 km, the upper bound placed on D. P2 prefers a depth of 18 km. The best-fit depths of
 287 all profiles are within uncertainty of each other. The total relative velocity across the
 288 shear zone is 7.2 ± 0.3 mm/yr, 6.8 ± 0.2 mm/yr, 8.4 ± 0.2 mm/yr, and 10.1 ± 0.2 mm/yr
 289 for P1, P2, P3, and P4 respectively.

290 6 Discussion

291 The favorable fit of the SZM to the geodetic profiles suggests that the deformation
 292 of the lower crust in the WL region is characterized by distributed viscous shear. How-
 293 ever, the observed geodetic strain can also be explained by the combined effect of EDM-
 294 related deformation across multiple faults, if the slip is roughly equally distributed among
 295 the dislocations. The presence of the near-vertical strike-slip faults in the northern sec-
 296 tion of the WL allows us to explore this possibility.

297 We show that both models fit the data quite well, as indicated by the data mis-
 298 fit. We cannot say whether one model fits better than the other, since direct statistic
 299 comparison of EDM and SZM is not possible since profile P1a is a subset of profile P1.
 300 However, due to the linear nature of the SZM, we can make a direct comparison between
 301 the EDM fit and a linear model fit to the same profile P1a.

302 6.1 Elastic Dislocation Model

303 The linear fit to profile P1a (Figure 6a) yields an RMS value of 0.23 mm/yr, which
 304 is essentially same as the RMS of the EDM (0.22 mm/yr). If the deformation is accom-
 305 modated by elastic dislocations, there are several possible reasons for the lack of a clear
 306 preference for the EDM over the linear model: (1) the noise in the data obscures the EDM,
 307 (2) the modeled fault locations or locking depths do not correspond to reality, (3) there
 308 are unknown dislocations present, and (4) the faults are late in their seismic cycles.

309 6.1.1 *Is EDM Hiding in the Noise?*

310 To address hypothesis (1), we estimate the likelihood of the observed result, assum-
 311 ing the presence of an EDM in the data. To that end, we construct synthetic data us-
 312 ing Equation 1 with fault parameters identical to those in profile P1a. We then gener-
 313 ate 100K realizations of noisy synthetic data by adding noise to the predictions of the
 314 model. The noise is randomly chosen from a normal distribution with a mean of zero and
 315 a standard deviation of 0.22 mm/yr, the estimated level of uncertainty in our GPS ve-
 316 locity data. We fit the line and the EDM to each of the noisy synthetic datasets and count
 317 how often the RMS of the EDM is smaller than that of the line by more than 0.01 mm/yr.
 318 Our findings indicate that there is a 78% likelihood of us being able to recover the EDM
 319 from the synthetic noisy data. This suggests that it is unlikely that the signal of EDM
 320 faults is obscured by the noise in the real data.

321 6.1.2 *Are There Better Fault Locations or Locking Depths?*

322 Regarding hypothesis (2), we note that the WL is an immature fault zone char-
 323 acterized by a complex geometry. For the purposes of modeling, the faults are represented
 324 as single straight lines, so there is a margin of error to the locations of the faults in the
 325 profile. The locking depths of each fault are similarly uncertain. To address the location
 326 uncertainty, we search for alternative fault locations within the profile that would result
 327 in a better fit of the EDM to the data. We test possible fault configurations by allow-
 328 ing each fault to vary its location between the nearest profile edge and the midpoint of
 329 the profile. We then fit each configuration with the EDM and search for a reduction in

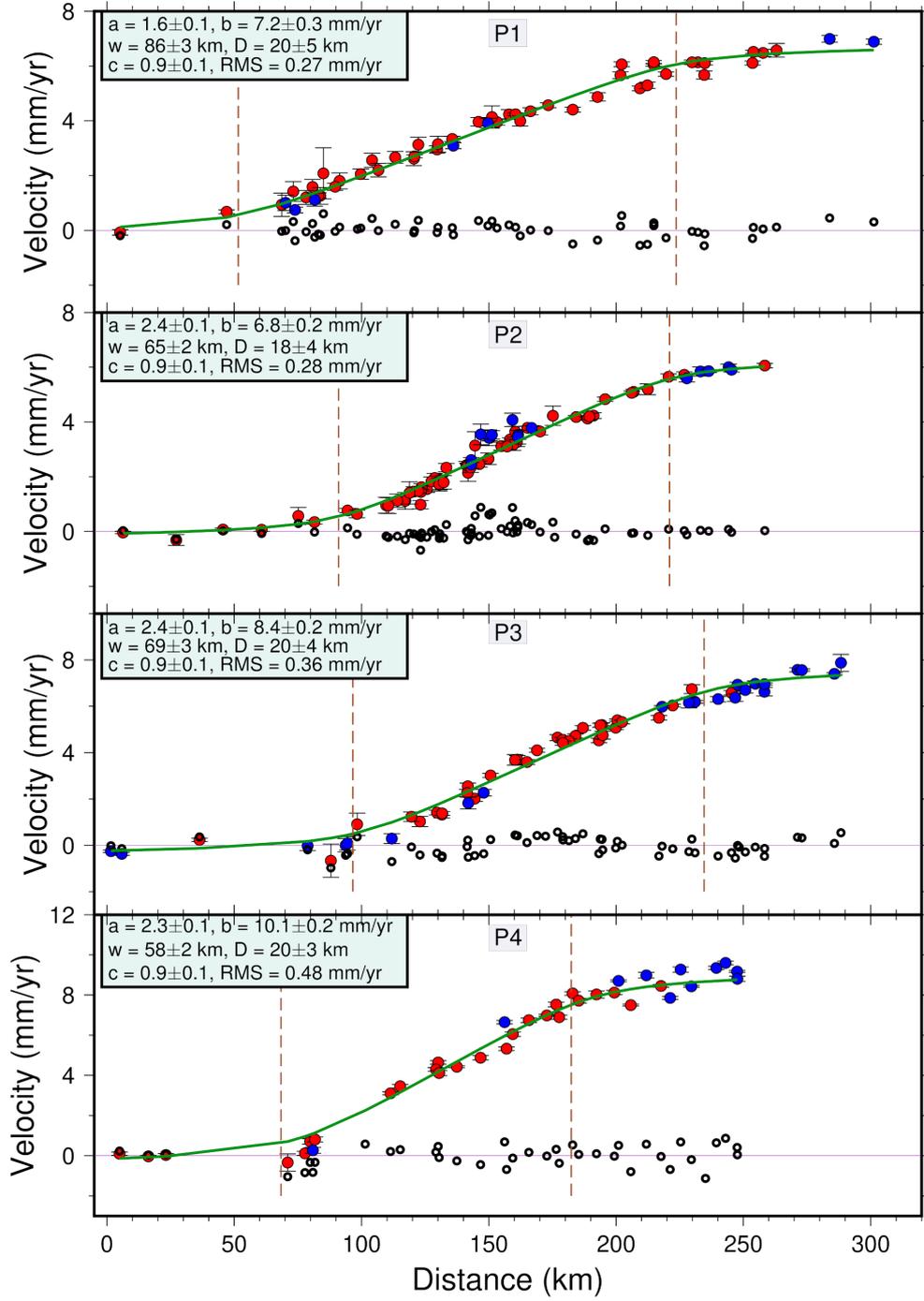


Figure 7. Shear Zone Model fits (Equation 2) to the across-profile velocity components are shown (the sign of the velocities is flipped). Station dots are color coded as in Figure 1d based on velocity azimuth. Best-fit model parameters are listed for each profile: horizontal shift factor a , velocity difference across the zone b , half-width of zone w , thickness of the elastic layer D , and the vertical shift factor, c . Model-derived shear zone widths at depth D are shown in dashed vertical lines. Residuals to the fit are also shown (small black circles).

330 the RMS error. We employ a similar approach to assess the uncertainty in the locking
331 depths of the faults, allowing the depths to vary between 2 and 30 km.

332 Our findings reveal that the fault locations that minimize the misfit are approx-
333 imately 19 kilometers for MV and 75 kilometers for HLWS. These locations are very close
334 to the assumed (3 km difference for MV and 5 km for HLWS), but do not appear to cor-
335 relate to seismicity clusters and result in an insignificant reduction of RMS by only 0.01
336 mm/yr. The preferred locking depths, which result in the same reduction in RMS as the
337 optimal locations, are approximately 15 kilometers for MV and 30 km for HLWS faults.
338 While 15 km locking depth for MV is feasible, the locking depth for the HLWS is un-
339 realistic, considering that relocated seismicity predicts a much shallower seismogenic thick-
340 ness of about 14 km.

341 An RMS contour plot for the locking depths of the two faults (Figure 8) indicates
342 that the locking depths are poorly constrained by the data, suggesting that using seis-
343 mogenic depths as a priori constraints may be more suitable for the analysis. The con-
344 tour plot for the locations of the faults shows that the location of MV is well constrained,
345 however there is much less preference for the location of the HLWS fault. The latter is
346 surprising, since the surface velocity in the EDM is driven by the location of the faults.
347 This is another feature of the data that is inconsistent with the downdip extension of
348 the HLWS.

349 *6.1.3 Are There Unknown Dislocations?*

350 Testing hypothesis (3), i.e., that there are additional dislocations present, presents
351 a challenge due to the limitations imposed by the relatively short length of the profile.
352 Given that the seismogenic thickness in the region varies between approximately 10 to
353 20 km, the critical length, πD , is between 30 and 63 km. The length of profile P1a is 115
354 km, which is 1.9 to 3.8 times the critical length, meaning that we may be able to resolve
355 a third dislocation, but no more.

356 We find that, despite a preference, the model's fit is not highly sensitive to the lo-
357 cation of the third dislocation: the difference between the minimum and maximum RMS
358 values is less than 0.02 mm/yr. If we begin the search with the initially assumed loca-
359 tions of MV and HLWS, the predicted location for the third fault falls between the west-
360 ern edge and 21 km or between 57 km and the eastern edge of the profile, with an RMS
361 value of 0.22 mm/yr (Figure 9, brown line). If we initiate the search with the best-fit
362 locations of MV and HLWS, as described in the previous paragraph, the predicted lo-
363 cation range of the third fault is similar, with an RMS of 0.21 mm/yr (Figure 9, orange
364 line). Any location within these ranges provides an equally good fit, but the reduction
365 in the RMS is essentially zero ($\Delta\text{RMS} < 0.001$ mm/yr). There is evidence for the ex-
366 istence of a strike-slip zone (Pyramid Lake fault zone, Eisses et al., 2015) in the eastern-
367 most section of this location window. However, the result of our analysis suggests either
368 that we are unable to resolve dislocations beyond the two already considered, or that ad-
369 ditional dislocations are not necessary, as they do not significantly enhance the fit of the
370 EDM.

371 *6.1.4 Are the Faults Late in their Seismic Cycle?*

372 Concerning hypothesis (4), several authors (e.g. Wang et al., 2021) point out that
373 the effects of coupling between the brittle upper crust and underlying viscoelastic lay-
374 ers on the earthquake cycle (Savage, 2000) cannot be ignored. The rate of strain accu-
375 mulation slows with time since last earthquake t , making the surface velocity appear more
376 linear late in the seismic cycle, i.e., if the time since the last event is significantly longer
377 than the relaxation time τ . Geological studies of past seismic events show that the MV
378 (Gold et al., 2014) is about mid-cycle, HL (Wills & Borchardt, 1993) is mid-cycle or less

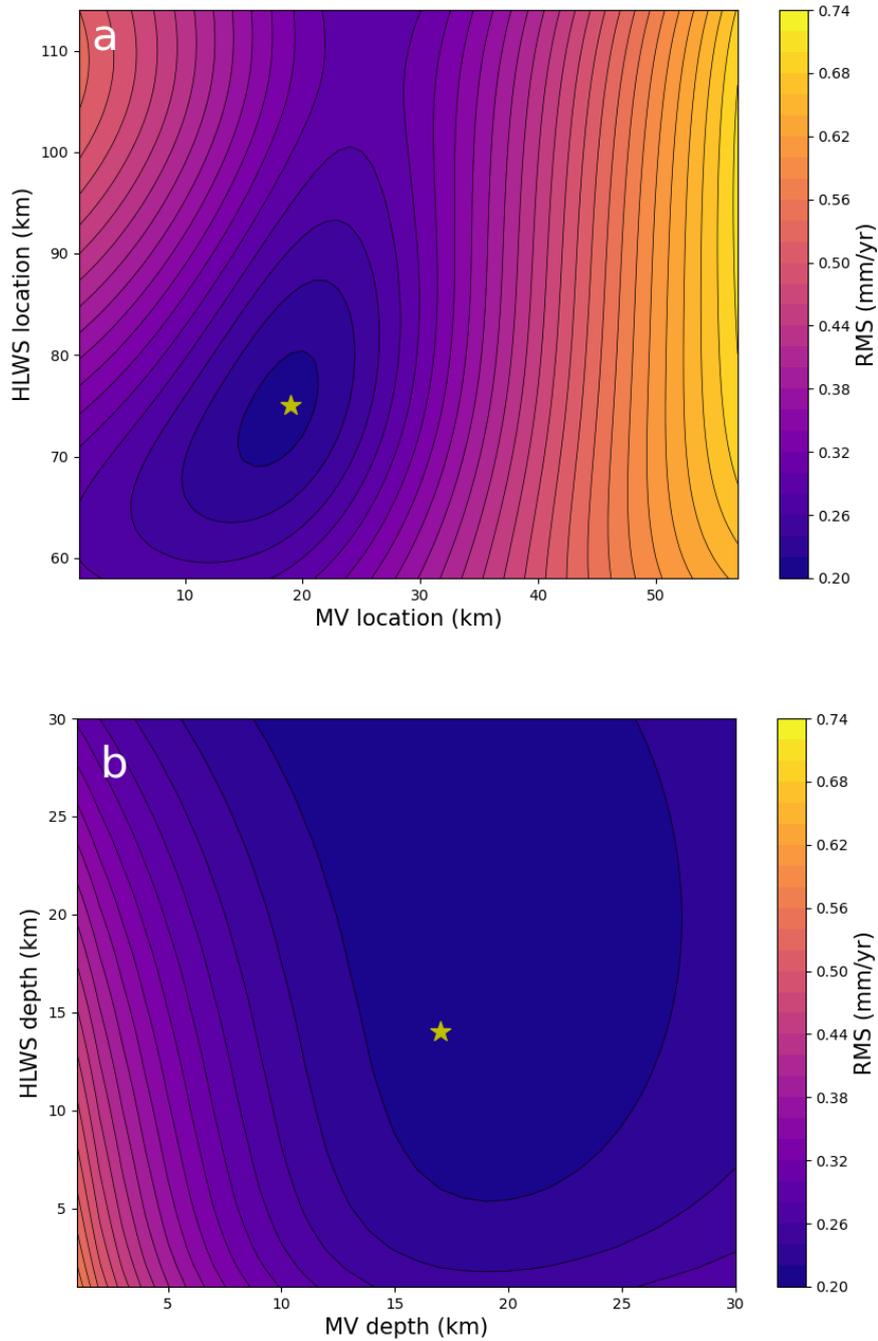


Figure 8. RMS contour plots for the locations (top) and locking depths (bottom) of the two modeled faults in profile P1a (i.e., HLWS = Honey Lake and Warm Springs faults, and MV = Mohawk Valley). The gold stars denote the best-fit locations and assumed depths for the purposes of fitting the EDM.

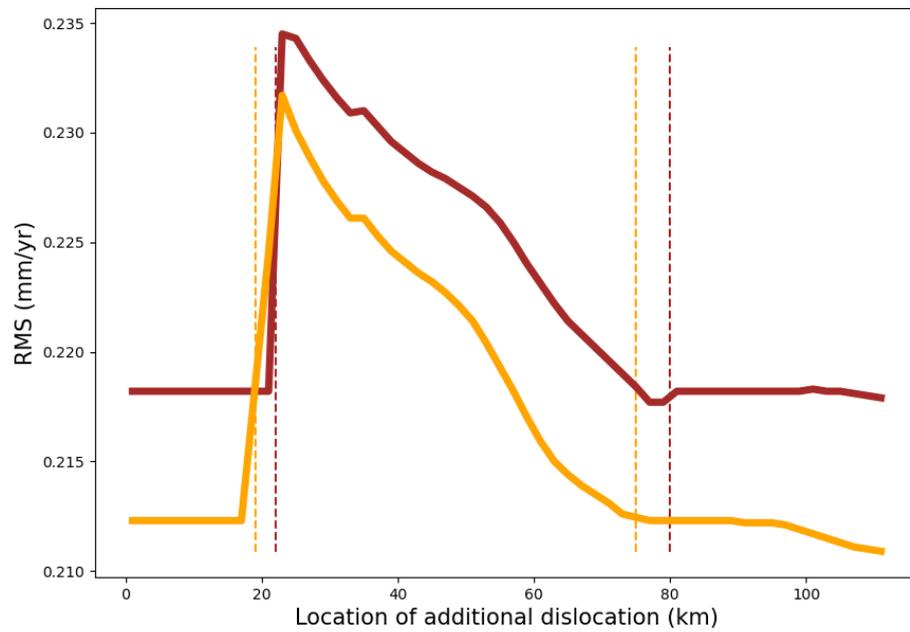


Figure 9. RMS misfit from a three-fault Elastic Dislocation Model as a function of the location of the third fault. Brown line is starting with assumed a priori locations of MV and HLWS faults, orange – best-fit locations (section 6.1.2). The locations of the MV and HLWS faults are shown in orange (best-fit) and brown (assumed a priori) dashed lines. Note that the difference between minimum and maximum RMS is less than 0.02 mm/yr.

379 (latest seismic event within a few hundreds of years), and WS (Chupik et al., 2022) is
 380 as early as ~ 100 years into its cycle. Taking a typically-reported viscosity for the lower
 381 crust, 10^{20} Pa s (e.g. Bills et al., 2007; Hammond et al., 2009), the ratio of the earth-
 382 quake recurrence time T to τ is longer than 15 for any of the faults in the northern WL.
 383 These values may be large enough to cause the velocities across the faults look linear,
 384 however the faults are in different stages of their seismic cycles, so it is reasonable to ex-
 385 pect differences in the slope of the velocity profile across them, which we do not observe.
 386 Furthermore, we have profiles P2, P3, and P4, which provide more opportunities for the
 387 faults within them to exhibit that they are in different times of their relaxation cycles.
 388 However, all profiles show similar linearity, increasing the likelihood that we are not de-
 389 tecting time since last seismic event with GPS velocities.

390 **6.2 Shear Zone Model**

391 We examined potential reasons for the lack of a clear preference for the EDM in
 392 profile P1 and found no compelling explanations. Despite different faulting styles cap-
 393 tured by P2, P3, and P4, their velocity profiles have similar characteristics as in P1. We
 394 will now present arguments supporting the presence of a distributed shear zone in the
 395 lower crust.

396 **6.2.1 Support for Distributed Shear Zone**

397 Distributed seismicity in the region (Figure 10) supports the idea of distributed de-
 398 formation of the lower crust. Specifically, with the exception of the MV fault, seismic-
 399 ity within profile P1a does not seem to correlate strongly with known faults. The dif-
 400 fuse seismicity implies that the lower crust within the WL is deforming more evenly than
 401 would be predicted by focused deep dislocations which transfer stress upward into the
 402 upper crust.

403 Lack of dislocation continuation into the ductile portion of the crust is also sup-
 404 ported by seismic imaging of the Warm Springs fault zone (Briggs et al., 2021), which
 405 revealed that the fault sections truncate at a depth of 8-12 km, intersected by a mid-crustal
 406 low-angle fault. Furthermore, uniformly featureless character of the velocity profiles is
 407 a strong argument for distributed shear. Despite significant geological variations and dif-
 408 fering faulting styles from north to south and from west to east, all four velocity pro-
 409 files exhibit the same shape. This similarity suggests that the same mechanism is respon-
 410 sible for surface deformation in all of these profiles.

411 The presence of a distributed shear zone in the lower crust aligns with the obser-
 412 vations of the surface features in the central WL. The presence of en echelon basins and
 413 rotated crustal blocks has been shown to be consistent with a uniformly shearing vis-
 414 coelastic layer, capped by the brittle upper crust (Wesnousky et al., 2012). The rotat-
 415 ing crustal blocks can be interpreted as rigid blocks riding on the underlying viscoelas-
 416 tic layer (Prescott & Nur, 1981; Wesnousky, 2005a).

417 **6.2.2 Summary of Arguments for Distributed Lower Crustal Shear**

418 Several factors support the SZM as the correct model:

- 419 1. The velocity profiles appear linear and the data is well-explained by the SZM (Fig-
 420 ure 6).
- 421 2. The SZM-predicted shear zone width aligns with that derived from the velocity
 422 azimuths (Figures 6 and 10).
- 423 3. The SZM-predicted bounds of the shear zone coincide with the edges of seismic-
 424 ity in the region (Figure 10).

- 425 4. The velocity profiles appear to be independent of fault geometry: it is impossi-
 426 ble to determine fault locations within each profile without prior knowledge of their
 427 locations.
- 428 5. Despite diverse surface features, all four profiles exhibit spatially consistent ve-
 429 locity profile shapes, suggesting a common deformation mechanism.
- 430 6. The faults are in different stages of their seismic cycles, yet that is not reflected
 431 in the uniformly linear velocity profiles.
- 432 7. Distributed seismicity in the region supports the idea of distributed deformation
 433 of the lower crust.

434 **6.3 Implications**

435 **6.3.1 Tectonics**

436 The observed distributed shear supports the conclusions of other studies that sug-
 437 gest the WL is a structurally immature plate boundary (Faulds et al., 2005; Wesnousky,
 438 2005b). Norris and Toy (2014) suggest a model for transform fault evolution, in which
 439 major continental boundary transforms begin as zones of broadly distributed shear, char-
 440 acterized by a number of smaller faults. These faults are limited to the seismogenic crust,
 441 eventually propagating into the lower crust and upper mantle due to positive feedback
 442 loops as the transform boundary matures. In this view, the dislocations in the WL will
 443 eventually organize into straighter and longer transform faults, which may develop lo-
 444 calized shear in the ductile lower crust and upper mantle.

445 Distance along the WL can be considered a proxy for the geological time, with the
 446 northern section being the youngest. Our findings show southward narrowing of the shear
 447 zone, indicating that the WL is becoming narrower over time. This supports the idea
 448 of the WL transforming into an incipient transform boundary.

449 We identified a distributed shear zone as the cause of surface deformation in the
 450 WL. However, we cannot distinguish whether the system is being driven from the sides
 451 (i.e. the Sierra Nevada block motion relative to the Basin and Range) or from below (Savage,
 452 2000). Barbot (2020) shows that the lower crust/upper mantle flow in the northern and
 453 central WL is uniform, and presents evidence that the brittle crust is mechanically cou-
 454 pled to the ductile portion of the lithosphere. This implies that the surface deformation
 455 in the WL, as well as the rotation of the Sierra Nevada, are ultimately driven by the deep
 456 interaction between the Pacific and North American plates.

457 **6.3.2 Consistency with Other Observations Suggesting Lower Crustal** 458 **Flow**

459 The absence of dislocations in the lower crust is consistent with ductile flow since
 460 substantial flow would inhibit the progressive development of stable planar zones of shear.
 461 Moreover, the time scale of flow observed in the lower crust tends to be shorter than earth-
 462 quake recurrence times, suggesting weakness of the lower crust on geologic time scales.
 463 The time scales of flow are indicated by studies of postseismic relaxation and isostatic
 464 rebound (Bills et al., 2007; Freed et al., 2007; Hammond et al., 2009; Dickinson et al.,
 465 2016), which estimate viscosities of about 10^{20} Pa s. This value implies that the relax-
 466 ation time for the lower crust is on the order of hundreds of years, which is short com-
 467 pared to the earthquake recurrence intervals of thousands of years. This is also supported
 468 by seismic reflection observations of Moho topography which is different from the sur-
 469 face topography (Hauge et al., 1987; McKenzie et al., 2000), as well as a very thin ap-
 470 parent elastic plate thickness in the Basin and Range estimated from the lack of coher-
 471 ence between gravity and topography (Lowry & Smith, 1994).

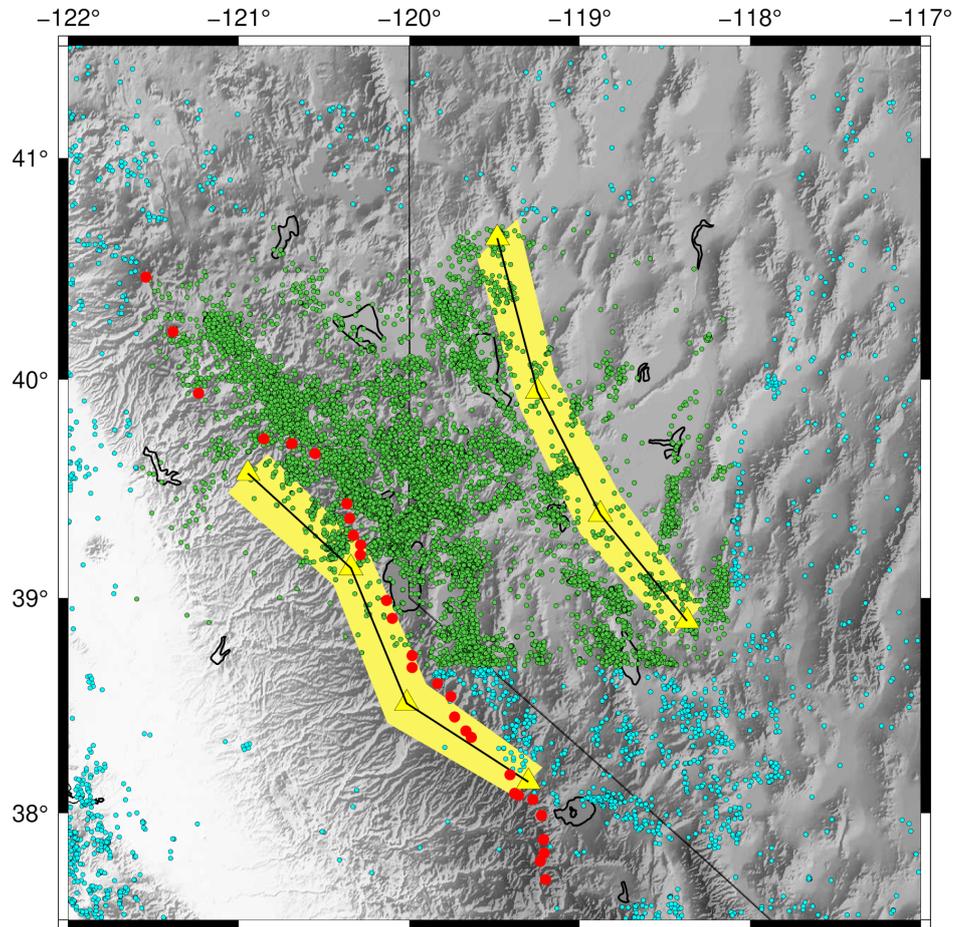


Figure 10. The map from Figure 1 is shown with the shear zone outline (black lines), predicted by the Shear Zone Model. Yellow shaded areas represent the uncertainty of the shear zone boundaries. Yellow triangles denote the location of the edges of the shear zone below the elastic layer of the crust in each profile. Relocated seismicity (Ruhl et al., 2020) with $m \geq -1$ between 2002-2019 is plotted in green where available, ANSS/ComCat mainshocks with $m \geq 2$ between 2003-2020 are plotted in blue elsewhere. Red dots are the tallest regional peaks, plotted as a proxy for the Sierra Nevada crest.

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6.3.3 Fault Slip Rates

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Our findings raise questions about the possibility of estimating slip rates on individual faults using geodetic data in areas where the surface velocities reflect distributed shear deformation. Bourne et al. (1998) suggested that the slip rates of individual faults are controlled by their number and location within the shear zone, not by deep dislocations beneath the faults. In this view, the slip rates are obtained by distributing the far field velocity budget onto the faults within the velocity profile. The slip rate on a given strike-slip fault is the difference between average velocities on either side of the fault. However, in the WL, this method can only be applied to the MV and, perhaps, the GV faults, since the shear zone extends horizontally past the strike-slip faults into an area characterized by normal faulting. Following the Bourne et al. (1998) approach, the slip rate for the MV fault is 0.72 mm/yr, which is within the range of permissible geologic rates with a minimum 0.4 mm/yr (Sawyer et al., 2013). Including the GV fault yields slip rates of 0.28 mm/yr for MV and 0.52 mm/yr for GV. The only geologic slip rate available for the GV fault is the general estimate of <1 mm/yr (Gold et al., 2013), with which our estimate agrees. However, that would imply that the slip rate for the MV fault is smaller than the minimum inferred geologic estimate.

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The SZM predicts that the total upper crustal fault slip rates across the entire shear zone should agree with geodetic estimates of far field motion, about 7 mm/yr to 10 mm/yr (depending on profile P1-P4). The sum of maximum geologic slip rates (rate plus uncertainty) on documented faults captured by profile P1a (Gold et al., 2013; Sawyer et al., 2013; Gold et al., 2014, 2017; S. Angster et al., 2016) aligns with the geodetic velocity budget across profile P1a of approximately 5 mm/yr. This leaves at least 2 mm/yr to be accommodated by the normal faults in eastern part of P1, however it is unclear how they accommodate the shear.

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A discrepancy between the total geodetic slip and that obtained by summing the geologic slip rates on the known faults has been noted (Hammond et al., 2011; Gold et al., 2014; Bormann et al., 2016; S. J. Angster et al., 2019). In the central WL, the missing geological slip has been attributed to the block rotations and strike-slip faults that may be missing from the geologic datasets (Dong et al., 2014; Bormann et al., 2016; Pierce et al., 2021). It is probable that processes like those occurring in the central WL may also be in effect in the northern WL. However, the vertical axis block rotations in the central WL are partly accommodated by east-northeast striking sinistral faults (Wesnousky, 2005a; Wesnousky et al., 2012; DeLano et al., 2019), which are not present in the northernmost WL. The lack of vertical axis rotations in the northern WL simplifies the estimation of the SZM-geodetic slip rates and makes comparison with geologic rates more straightforward. In any case, using the SZM-based geodetic slip rates does not in and of itself explain the discrepancy between geologic and geodetic slip rates, nor does it make the discrepancy worse. It does, however, change some of the details in the geodetic slip rates estimates and could lead to somewhat different estimates of seismic hazard distribution, if it is based on SZM-based geodetic slip rates.

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7 Conclusion

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Our study challenges the use of elastic dislocations with deep creep for explaining active deformation along faults everywhere. Geodetic evidence in the northern and central Walker Lane supports a distributed shear zone in the lower crust. This suggests that faults likely terminate near the bottom of the upper crust. Consequently, models based on discrete dislocations in the viscoelastic lower crust are not appropriate to estimate the slip rates on the individual faults. A more suitable approach in these locations is to consider how the total relative velocity budget is distributed among the faults, potentially based on the location and azimuth of the fault within the shear zone.

8 Open Research

All data used in this research can be accessed freely on the Nevada Geodetic Laboratory website (<http://geodesy.unr.edu>). The GPS data has been collected through the MAGNET GPS network, EarthScope Network of the Americas (Community, 2006), the Washoe County GPS Network, and Leica SmartNet Network. Maps and figures were created with Matplotlib version 3.7 (Hunter, 2007) (<https://matplotlib.org/>) and Generic Mapping Tools (GMT) version 6 (Wessel et al., 2019) (<https://www.genericmapping-tools.org/>).

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