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1	Persistent fine-scale fault structure and rupture development: a new twist in the
2	Parkfield, California, story.
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14	ABSTRACT
15	We investigate the fine-scale geometry and structure of the San Andreas Fault near
16	Parkfield, CA, and their role in the development of the 1966 and 2004 \sim M6 earthquakes.
17	Long-term surface fault traces indicate that structural heterogeneities associated with
18	secondary reverse and normal fault structures are present at both rupture tips, near
19	Middle Mountain and Gold Hill. Detailed analysis of almost 50 years of high-resolution
20	seismicity reveals a fault plane that has been twisted into a helicoid between Middle
21	Mountain and Gold Hill. Numerical models support our conclusion that this shape is the
22	result of long-term torqueing of a strong stuck patch surrounded by a weak creeping
23	region. The changes in fault friction behavior and related geometric discontinuities act
24	as barriers to rupture propagation of moderate size earthquakes at Parkfield, and as
25	areas of concentrations where rupture initiates. Our study demonstrates also that

smooth strike-slip faults with large cumulative offset can form new fault segments at alate stage in their evolution.

28

29 1. Introduction

30 The Parkfield section of the San Andreas fault is one of the best-instrumented 31 earthquake source region in the world. Over the past 50 years, one and a half seismic 32 cycles, including two M~6 earthquakes in 1966 and 2004, have been observed there 33 with full geodetic and seismic coverage. Yet it is still unclear how the earthquake 34 generation process takes place in this peculiar area. It has long been recognized that the 35 M~6 earthquakes break a stuck patch imbedded in an otherwise creeping fault – first 36 noted by Harris and Segall (1987) and most recently modeled by Barbot et al. (2012). It 37 has also been observed that the rupture zone lies between two geometrical 38 discontinuities in the fault, a slight compressional bend in the north near Middle 39 Mountain (MM) and an extensional jog in the south near Gold Hill (GH) (Lindh and 40 Boore, 1981). Neither the connection between these features nor their role in the 41 nucleation and termination of the $M \sim 6$ earthquake ruptures have been elucidated.

42 The 1966 earthquake initiated at the northern end of the rupture area near Middle 43 Mountain, propagated unilaterally to the southeast along the entire Parkfield section 44 and stopped near Cholame (CH) (Fig. 1 and 2; Brown et al., 1967; McEvilly et al., 1967). 45 In contrast, the 2004 earthquake initiated in the south near Gold-Hill (GH), then propagated mainly to the northwest and stopped near MM (Fig. 1 and 2; Langbein et al., 46 47 2005; Rymer et al., 2006). Both surface ruptures were similar, formed by small fractures 48 associated with local displacements less than 0.1 m (Brown et al., 1967; Rymer et al., 49 2006). Source inversion models of the 2004 event reveal two main slip patches close to 50 either end of the rupture extent, with maximum slip ranging from 0.5 to 1m (see an example in Fig. 3), followed by post-seismic afterslip near the surface above the
coseismic rupture zone (e.g., Barbot et al., 2009; Johanson et al., 2006; Langbein et al.,
2006; Murray and Langbein, 2006) but also at depth in the lower crust (Bruhat et al.,
2011). Similar co-seismic distribution associated with shallow postseismic behavior has
been observed for the 1966 earthquake (see an example in Fig. 3; e.g. Archuleta and Day,
1980; Scholz et al., 1969; Segall and Du, 1993).

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58 In order to better understand rupture initiation, propagation and arrest of the two latest 59 M6 earthquakes at Parkfield, we present here a combined analysis of high-resolution 60 near surface and earthquake data that image the fine-scale structures of the San Andreas 61 Fault. We bring together geologic fault information available in the literature and a 62 broad range of satellite images and topographic data to refine the long-term surface 63 fault traces along the Parkfield section. We also compute high-resolution seismicity 64 recorded at the Northern California Seismic Network (NCSN) between 1969 and 2015, 65 and aftershocks recorded during the 2 months following the 1966 earthquake (Eaton et 66 al., 1970), spanning a nearly 50 year time period including two ~M6 events. Using these 67 new data, we relate the structures observed at the surface to persistent deformation 68 patterns along the three-dimensional fault plane at depth. Supported by numerical 69 simulations of the long-term fault deformation, we propose a new mechanical model for 70 the Parkfield segment of the San Andreas fault that differs from previous studies (e.g., 71 Simpson et al., 2006).

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73 2. Data analysis and results

74 2.1 Surface fault traces

75 According to the quaternary surface fault map from the USGS (U.S. Geological Survey 76 and California Geological Survey, 2006); blue lines Fig.1b and 1c), the \sim 30 km long 77 Parkfield section is composed of two linear, roughly N139°E striking, sub-parallel 78 strands (also called the main San Andreas fault and the southwest fracture zone) that 79 join each other at both terminations of the Parkfield section through two structural 80 complexities : i) at the southern end near GH, an extensional jog, where the fault trace 81 bends twice with deviations up to 28° from the mean fault strike (GH-CH, Fig.1b and 2c); ii) at the northern end, near MM, a 5° restraining bend (Lindh and Boore, 1981) (MM, 82 83 Fig.1b and 2c).

We analyzed multi-resolution satellite images (Google Earth, Landsat 7) and 84 topographic data (SRTM 30 m, LiDAR available from opentopography.org) and identified 85 86 additional structural features (black lines in Fig. 1b and 1c): around MM, the active fault 87 trace is surrounded by multiple secondary faults several kilometers away from the main 88 San Andreas Fault and clearly imprinted in the topography (Fig. 1c). Most of them have 89 been previously recognized as secondary strike-slip and reverse faults in the field 90 (Thayer, 2006). Altogether they contribute to forming the localized relief observed at 91 the center of the main fault trace. The overall shape of this structural feature, which 92 encompasses an area approximately 5 km wide and 10 km long, resembles a restraining 93 step-over feature that separates the locked Parkfield section from the creeping section. 94 These secondary faults strike between N110°E and N150°E, in good agreement with 95 earthquake locations derived in this study and focal mechanisms from Hardebeck 96 (2010) that show oblique reverse and normal faulting (Fig. 1c; see also Thurber et al., 97 2006). The 1966 and 2004 earthquakes initiated and stopped near these large structural 98 discontinuities observed at the surface near GH-CH and MM.

100 *2.2. Earthquake relocation*

101 We use seismicity data to investigate the depth extent of the heterogeneities observed 102 near the surface. We first apply a 3D grid search method (NLLoc program; Lomax et al., 103 2000) to determine new absolute locations of more than \sim 17,000 earthquakes from 104 1966-2015 in the 3D velocity model of Thurber et al. (2006). The strike of the new 3D 105 absolute locations is rotated slightly (up to $\sim 2^{\circ}$) counterclockwise compared to the 106 original NCSN locations as a result of properly accounting for the velocity variations 107 across the fault zone. We then use the double-difference algorithm HypoDD 108 (Waldhauser, 2001; Waldhauser and Ellsworth, 2000) to compute precise relative 109 locations using both travel-time differences from P-wave pick data (available from 1966 to 2015) and P- and S-wave cross correlation delay time data (1984-2015) (Schaff and 110 111 Waldhauser, 2005). We included all events with at least 4 stations in order to be able to 112 simultaneously relocate the aftershocks of the 1966 earthquake (Eaton et al., 1970) 113 together with the rest of the seismicity. The resulting locations are shown in Fig 1, S1c 114 and S1d. These locations resolve the fine shape of the fault plane at a level of a few tens 115 of meters. The median lateral relative error at the 95% confidence level, derived from a 116 bootstrap analysis of the fine residuals, is 0.039 km and the median vertical relative 117 error is 0.022 km (Fig. S2).

118 Most events with M < 1 (M_L is considered for events M > 3, M_D for smaller events; see 119 Uhrhammer et al., 2011) are located in the creeping section (see Supplementary Fig. S1a 120 and S1b). A cross-section at MM (Fig. 1d) indicates that some events also delineate 121 secondary oblique strands in the eastern part (see black arrows) that connect to the 122 main fault at ~10 and ~12 km depth, close to the 1966 nucleation area. Several reverse 123 and normal faulting focal mechanisms located between 10 and 15 km deep (events 3, 4, 124 5 and 6 in Fig. 1c and 1d) are aligned with those secondary strands, while those located between 0 and 10 km deep (events 1, 2 and 7 in Fig. 1c and 1d), appear to be related tothe secondary fault traces observed at the surface (black crosses, Fig. 1d).

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2.2.1. Distribution of the 1966 and 2004 aftershocks

129 In the following analysis, we only consider earthquakes of $M \ge 1$, corresponding to the 130 minimum magnitude of completeness of the earthquake catalog (see Supplementary Fig. 131 S3). Fig. 2 shows the distribution of the 1966 and 2004 aftershocks within 2 months of 132 the mainshock (in blue and red, respectively). The along strike distribution of the 133 aftershocks is sharp and fairly similar for both events, mostly situated between 4 and 12 134 km deep along the Parkfield section. Distinctive seismicity patches (e.g. streaks, clusters, 135 repeating events, Waldhauser et al., 2004; Thurber et al., 2006) outline an area devoid of 136 earthquakes between 6-9 km deep, corresponding to the main locked patch of the fault 137 that broke in the 1966 and 2004 events (dashed lines in Fig. 2b, Harris and Segall, 1987). 138 The 1966 and 2004 aftershock sequences are similarly distributed, especially near GH-139 CH where two distinct patches of seismicity, at ~ 5 km and ~ 10 km depth, are 140 superimposed (Fig. 2b), suggesting that the same main fault strand has been activated 141 during the two earthquakes. A high concentration of number of aftershocks (including 142 all $M \ge 4$ events) and cumulative moment release are observed near the rupture tips of 143 the 1966 and 2004 earthquakes (Fig. 2a, b; Fig. 3a, b).

We note that in the northwestern part of the Parkfield section, the 1966 aftershocks tend to deviate further to the northeast away from the San Andreas fault trace (blue dots in map view on Fig. 2a). This deviation is also observed in the original absolute locations and might be due to the poor station coverage in the northwestern part of the 1966 rupture zone (Eaton et al., 1970). However, few events are concerned and our relocations are in good agreement with previous studies in the southeastern part of the

Parkfield section, where most of the seismicity is localized (Fig.2; e.g., Thurber et al., 2006; Waldhauser and Schaff, 2008). More generally, less than 20 permanent and temporary stations recorded aftershock activity following the 1966 earthquake, with several periods of interrupted service (Eaton et al., 1970). This likely explains why the aftershock activity and the cumulative seismic moment release shown in Figure 3 are much lower for the 1966 sequence compared to the one in 2004.

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2.2.2. Interseismic periods 1969-2004 and 2005-2015

158 We choose to represent in Fig. 2 only the best (i.e., correlation based) relocations (gray 159 dots) for the two interseismic periods from 1984–2004 and 2005–2015 for a fine-scale 160 representation of the geometrical features (see Supplementary Fig. S1 for the entire relocated catalog). The interseismic seismicity distribution is remarkably similar to the 161 162 co- and postseismic distributions. Figure 3 shows along strike cross sections of the 163 earthquake density and cumulative seismic moment released within 2 months after the 164 1966 and 2004 Parkfield earthquakes (Fig. 3a and 3b, respectively), and during the 165 1984–2004 and 2005–2015 interseismic periods (Fig. 3c and 3d, respectively). Density 166 contour lines point out five, possibly six, persistent kilometer scale seismicity patches 167 during the interseismic periods (Fig. 3c and 3d). A patch of higher density is located at 168 the northern end of the Parkfield section at 4 km depth, a patch that was also activated 169 by aftershocks of the 2004 event (Fig. 3b).

The interseismic, cumulative moment release maps allow us to distinguish other patches
which are lining up at 4-6 km and ~10 km depth (Fig. 3c and 3d), again in good
agreement with patches that were active during the 1966 and 2004 post seismic periods
(Fig. 3a and 3b).

175 *2.2.3. 3D fault geometry*

176 We determine the three-dimensional fault geometry by applying a principal component 177 analysis to all events within 3 x 3 km wide boxes (in map view), stepping at 1 km 178 intervals along the fault trace. For each box, we calculate strike and dip of the plane that 179 minimizes the distance between hypocenters and fault surface during each of two 180 interseismic periods (1984-2004 and 2005-2015) in an attempt to capture the long-181 term signature of the fault geometry (Fig. 2c). For simplicity, we assume, in each box, a 182 constant dip of the calculated plane. The calculated strike and dip values are presented 183 in Fig. 2c, while an animated view of the data and the combination of the best fitting 184 planes is shown in Supplementary Fig. S6.

185 The strike measurements show that the strike along the mainshock rupture zone, which 186 includes the stuck patch, is fairly constant, while it deviates from that trend at both ends: 187 the strongest deviations range from N125°E to N150°E (at MM and GH), consistent with 188 surface measurements (see black line in Fig. 2c), albeit with fluctuating amplitudes. The 189 overall strike variations are also in good agreement with focal mechanisms derived by 190 Hardebeck (2010; Supplementary Fig. S7). Figure 2d shows the standard deviations of 191 the seismicity from the best fitting planes for each interseismic periods, characterizing 192 the normal distance of the events from the best fitting planes (see also Supplementary 193 Fig. S4 and S5 for detailed event distribution). It confirms that MM and GH-CH are 194 characterized by greater deviations of hypocenters from the plane (400-500 m), 195 suggesting a wider off-fault deformation zone, compared to the Parkfield section (100-196 200m). The systematic variation in strike along the fault indicates a twist in the fault 197 plane along a vertical axis.

Furthermore, the fault dip varies steadily from northeast dipping in the north tosouthwest dipping in the south, indicating a twist in the fault plane along a horizontal

axis (Fig. 2c; e.g. Kim et al., 2016). The 3D fault geometry of the San Andreas fault at
Parkfield can thus be best described as a helicoid (Supplementary Fig. S6). The change in
fault dip from northwest to southeast can also be observed in the original NCSN
locations, as well as the 3D grid-search absolute locations, indicating that the twist is not
an artifact of the relocation procedure or the model used to locate the events (see
Supplementary Fig. S8).

206 Because the seismicity patches are well defined and aligned, we used the principal 207 component analysis to minimize the orthogonal hypocenter distances to a single fault 208 plane in each box (Supplementary Fig. S4). However, we also tried to investigate the 209 possibility of multiple fault strands or variations of the fault dip at depth below and 210 above 7 km (i.e. the limit between the main patches of seismicity; see Supplementary 211 Fig. S4). The resulting fault planes were poorly constrained, especially the dip values, 212 because their fits were mostly controlled by single sub-horizontal streaks, rather than 213 events well distributed on a plane in three-dimensional space. Estimating an orientation 214 by fitting one single plane in each box tend to smooth local irregularities that might be 215 due to multiple fault strands at depth. Event distributions away from the fault planes 216 (Supplementary Fig. S5) show that the Parkfield section is mainly characterized by a 217 clear and sharp single strand of ~100m wide. Secondary peaks, indicating possible 218 oblique secondary strands, are observed around MM and GH-CH. Therefore, our 219 approach cannot describe in detail the complexity of the damage zone around the fault 220 core, but it allows us to highlight significant discontinuities with great confidence.

221

3. Numerical model of the effect of a strong stuck patch on fault geometry

The creeping section is known from studies at SAFOD to be very weak, with a friction coefficient $\mu \sim 0.15$ due to the presence of weak phyllosilicates in the fault zone (i.e. 225 saponite at shallower depth, Carpenter et al., 2015; Lockner et al., 2011, to talc below 3 226 km deep, Moore et al., 2016). They appear to be produced by reaction between the 227 serpentinite of the Coast Ranges ophiolite and the silicious rocks of the Salinian block 228 (Moore and Lockner, 2013) – a geologic context that is only present along the creeping 229 section (Allen, 1968; Irwin and Barnes, 1975). On the other hand, the stuck patch at 230 Parkfield cannot be lined with weak phyllosilicates because they are velocity 231 strengthening and will not facilitate seismic instability. The only possible materials that 232 can be velocity weakening and exhibit the frictional strength of the stuck patch in this 233 area are granite, antigorite or lizardite forms of serpentinite (Moore and Rymer, 2007). 234 These all have high strength with friction coefficients between 0.5-0.75. Thus, the 235 seismogenic stuck patch has a higher friction coefficient than the surrounding creeping segment and thus can sustain greater shear stress than the stress-drops observed in 236 237 Parkfield earthquakes (on the order of ~ 1 MPa). Since the earthquakes do not relieve all 238 the stresses on the stuck patch, there must be a permanent torque produced by that 239 high-stress patch surrounded by low stress regions. As a result, over time periods longer 240 than the seismic cycle, slow inelastic deformation will occur in response to this torque 241 and the twist in the fault plane will develop.

In the following section, we present a numerical model that supports the above idea. Rather than simulate the entire process of accumulating torque and inelastic deformations, we focus on showing that a high-friction patch embedded in a low-friction fault plane, representing the stuck patch situation, is able to generate deformations that are consistent with the proposed helicoidal twisting of the fault plane.

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248 *3.1 Model setup*

249 We use PyLith, a portable, scalable finite element software for simulation of crustal 250 deformation and earthquake faulting (Aagaard et al., 2016, 2013), to simulate the 251 deformation of linear elastic crust containing a straight strike-slip fault with 252 heterogeneous frictional strengths. The model domain, covering an area 40 × 162 km 253 wide and 21 km-thick (Fig. 4a), is discretized into hexahedral linear elements with edge 254 length of 1 to 1.5 km. Density of the crust is assumed to be 2500 kg/m³ and Vs and Vp to 255 be 3 and 5.2915 km/s. These values are equivalent to shear and bulk modulus of 22.5 256 and 55 GPa, respectively. Completely contained within the crust, the modeled zero-257 thickness fault plane is 90 km long and 20 km tall, going through the center of the 258 domain parallel to the longest edge of the domain. The creeping section is represented 259 by a low-friction section with a uniform friction coefficient of 0.1 and a cohesion of 2 260 MPa ("creeping fault plane" in Fig. 4a). The stuck patch is modeled as a simplified 261 rectangular zone of 20 × 5 km ("high friction patch" in Fig. 4; Harris and Segall, 1987). It 262 has a uniform friction coefficient of 0.6 and a cohesion of 2 MPa. The fault section that 263 broke during the 1857 earthquake ("high-friction fault plane" in Fig. 4), is considered to 264 have the same friction coefficient and cohesion as the stuck patch. All the fault sections 265 are allowed to slip because we model fault behaviors for a time period corresponding to 266 multiple earthquake cycles, over which no part of the Parkfield section would remain 267 completely locked.

Right-lateral motion at a rate of 2 cm/yr on two fault-parallel domain boundaries (relative on-fault slip rate of 4 cm/yr) with the normal component of velocity (v_x) set to be zero and the vertical component (v_z) free (Fig. 4b). On the fault-perpendicular sides, a simple-shear velocity field is applied such that the strike-parallel component varies between +2 and -2 cm/yr linearly with distance across the fault. The fault perpendicular component is set to be zero and the vertical component is free on these boundaries. The 274 bottom boundary is a free-slip surface. The top boundary is a free surface with zero 275 traction. The effect of gravity is considered and the initial stress is set to be lithostatic. 276 We get fault displacements and stresses as a response to incrementally increasing 277 amounts of boundary displacements. Although we apply velocities to generate strike-278 slip kinematics, the velocity boundary conditions are simply a way of increasing strike-279 slip motions incrementally at each time step. Our models simulate only linear elastic 280 behaviors of rocks and does not directly show rate-dependent effects. Nonetheless, the 281 correspondence principle (e.g., Sec. 2.5.7, LeMaitre, 2001) ensure the relevance to 282 permanent helicoidal twisting of the Parkfield section. According to this principle, strain 283 rate field in a linear viscoelastic model would be equivalent to elastic strain field in the 284 corresponding linear elastic model under the same boundary conditions. So, we can 285 view our model's strain field as a strain rate field for long term deformation.

286

287 3.2. *Results*

Figure 5 presents the northeastern displacements on the fault plane when the high-288 289 friction patch and the upper 2 km of the high-friction plane started slipping after 112 m 290 of cumulative displacement (Fig. 5a and 5c). After an unspecifiable number of 291 earthquake cycles, slip would accumulate on the high-friction patch as well. Thus, we 292 also look at the northeastern displacements when the entire high-friction patch as well 293 as the upper 5 km of the high-friction plane has slipped after 160 m of boundary 294 displacement (Fig. 5b and 5d). Relative to the state in Fig. 5a and 5c, northeastern 295 displacements have increased at the northern tip and more at the southern tip of the 296 patch. A major warp of the fault plane occurs at the northern tip of the high-friction 297 patch, corresponding to MM, with northeastern displacements > 1 m along a zone 298 approximately 10 km long (Fig. 5a and 5c). We note also that a section of the fault

situated in the southern tip of the stuck patch is slightly displaced toward the southwest
when most of the high-friction patch is locked (Fig. 5a and 5c), accentuating the general
warp of the fault plane. When the high friction patch starts to slip, the whole section is
progressively shifted toward the northeast, but the general warping of the fault plane is
still preserved (figure 5b and 5d). An animated view of the model is available in the
Supplementary material (see Supplementary Fig. S9).

305 Cross sections in Fig. 6 show the deformation of the San Andreas Fault plane at specific 306 locations along the model (Fig. 6a) after the stuck patch has been fully mobilized (i.e., 307 the stage in Fig. 5b and 5d). The vertical cross-section at MM (Fig. 6b) indicates a 308 northeastern displacement of the fault plane from its original location near the surface 309 (hence a southwest dipping fault). Most of the deformation diffused in the crustal 310 medium is fairly symmetrical relative to the fault plane. It seems to root at the base of 311 the stuck patch and to widen towards the surface in a flower-like structure. The 312 deformation is less pronounced at the southern tip of the stuck patch (i.e., GH, Fig. 6c). 313 Yet, we can distinguish a slight deflection of the fault plane, which is displaced toward 314 the northeast in the shallower and deeper part of the model (i.e. the creeping zones) and 315 less so in an area that corresponds to the stuck patch. Thus, the dip of the fault plane 316 changes with depth (i.e. toward the southwest, northeast and southwest). Figure 6d is a 317 horizontal cross-section situated at 7.5 km depth, crossing through the creeping zones 318 and the stuck patch (see Fig. 6a). It highlights the variation of the fault strike (grey 319 dotted line) compared to a referenced linear fault direction (black dotted line), in the 320 same way as in Fig. 2d. As shown in Fig. 5, the fault plane is mostly deformed toward the 321 northeast at MM and GH, and much less so in the southern half part of the stuck patch. 322 The strike variations within the stuck patch are consistent with the clockwise rotation of 323 the stuck patch inferred from the seismic data (Fig. 2c, 2d).

324 Figure 7 provides additional information on on-fault stress distribution at two stages: 325 after 112m of displacement (Fig. 7a), the maximum shear stress (~100 MPa) occurs at 326 the left (northwestern) tip and corners of the rectangular stuck patch (corresponding to 327 MM), which begins to slip. At this point, slip along the stuck patch, which is meant to 328 represent cumulative slip from repeated earthquakes, is about 30 cm with a stress drop 329 of ~ 1 MPa, which is regained in ~ 30 yrs. Yet, after 160 m of cumulative displacement, 330 stress remains at about 100 MPa throughout the seismic cycles, even after the high-331 friction patch has slipped (Fig. 7b). Thus, slow anelastic processes can produce 332 permanent deformation of the same type as the elastic ones shown in Fig 5. This result 333 indicates that the high-friction patch can uphold a higher level of stress than the 334 surrounding creeping section permanently, i.e., over a long time period corresponding to 335 many earthquake cycles. In both states, accumulating slip on the creeping section 336 generates high shear stress near the clamped bottom edge of the fault plane. Likewise, 337 shear stress is concentrated right beneath the slipping region on the high-friction section, southeast of the patch (Fig. 7b). Profiles of shear and normal stresses (σ_{xy} and 338 339 σ_{xx} , respectively) are shown in Supplementary Fig. S10. Along dip, the shear stress 340 increases on the stuck patch relative to the surrounding creeping regions and higher 341 stress concentrations are localized at the lower part of the stuck patch. Along strike, the 342 variation of shear and normal stresses are consistent with compression at the left tip 343 (MM) and extension at the right tip (GH-CH) of the stuck patch, resulting in clockwise 344 rotation.

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346 4. Discussion

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4.1.1. Shear stress and friction contrast at the origin of the twist

4.1. Progressive deformation of the San Andreas fault over the long-term.

349 Our observational results show that the San Andreas fault plane within the rupture 350 region of the recent ~M6 events has been deformed into a helicoidal surface that 351 extends from the compressional bend at MM in the north to the extensional jog near GH 352 in the south (Fig. 2c and Supplementary Fig. S6). It connects the northern part of the 353 fault to the Cholame section in the south through a ~ 2 km wide step. By incorporating 354 changes in fault-slip behavior at depth, our numerical model (Fig. 6) reproduces this 355 twist and thus the variation of the fault strike (Fig. 2d, 6d) and fault dip (Fig. 2d, 6b, 6c) 356 deduced from the seismicity analysis. The stuck patch is stronger than the surrounding 357 creeping areas by a factor of \sim 4 (see section 3). The contrasting high right-lateral shear 358 stresses on the patch will induce a torque that will bend the stuck section of the fault 359 clockwise, as shown in Fig. 8a (map view). Because the shear stresses increase with 360 depth, another torque increases the clockwise bending with depth (Fig. 8b). In our 361 numerical model, the increase in shear stress from the surface to 10 km depth is about 362 30 MPa in the creeping section away from the stuck patch (Fig. 7a, b). However, a much 363 stronger torque is resulted, about 120 MPa over the same vertical distance, through the 364 stuck patch (Fig. 7b). This result shows how the assumed contrast in frictional strength 365 can be responsible for the torques localized around the stuck patch. The two 366 mechanisms in Fig. 8a and Fig. 8b would produce the helicoidal fault geometry that we 367 observe (Fig. 8c). This deformed section acts like a hinge and gate, with the hinge in the 368 north marked by the compressional bend at its onset (MM in pink, fig. 8c), and the open 369 end of the gate to the south, where it is accommodated by the extensional jog (GH-CH in 370 blue, fig. 8c). This suggests that the geometrical complexities at the ends of the stuck patch result from the contrast of strength of the stuck patch with the surrounding 371 372 creeping section of the fault.

373

375 structures.

376 One hypothesis would consider that the geometrical complexities are inherited 377 structures formed before the presence of the stuck patch in the geologic evolution of the 378 San Andreas fault. They would correspond to ancient relay zones between fault 379 segments, where many WNW-ESE long-term secondary structures (Fig. 1c) would be 380 connected to the main San Andreas Fault. As such they could play an important role in 381 the fault structure and the rupture development of ~M6 Parkfield earthquakes. They 382 would contribute to control the extent of the rupture, acting as structural barriers (e.g., 383 Biasi and Wesnousky, 2017; Manighetti et al., 2007; Stirling et al., 1996) and places 384 where stress concentrations might also favor rupture initiation (e.g., Aki, 1979; Barka 385 and Kadinsky-Cade, 1988; Manighetti et al., 2015; Shaw, 2006). The secondary 386 structures are possibly oblique thrusts and folds formed in a transpressive regime 387 during the southern propagation of the San Andreas Fault (e.g. Perrin et al., 2016; Titus 388 et al., 2011 and references therein). Their obliquity varies depending on the distance 389 from the main San Andreas Fault (Titus et al., 2011) and some of them could be still 390 potentially active (ex: the 1983, Mw 6.2 Coalinga thrust earthquake, Stein and King, 391 1984 and 1985, Mw 6.1 Kettleman Hills thrust earthquake, Ekström et al., 1992). The 392 diffuse seismicity around Middle Mountain (Fig. 1, 2, S1, S4 and S5) might be explained 393 by the presence of these secondary faults in the crustal medium surrounding the main 394 San Andreas Fault. Yet, it seems unlikely that the presence of the stuck patch randomly 395 coincides or results from the presence of the structural complexities. The San Andreas 396 fault is a large cumulative slip fault with a very smooth (i.e., linear) surface fault trace 397 along the central section (Wesnousky, 1988), except at the tips of the Parkfield section. 398 Thus, it is necessary to find another mechanism responsible for the local twist of the

399 fault plane. We propose a second, more likely, hypothesis: the local structural 400 complexities were lately formed by the torque of the fault plane due to the presence of a 401 locked fault patch surrounded by creeping zones at depth. Thus, secondary geological 402 fault traces at MM and GH-CH should be considered in the common view of the active 403 fault traces in California. They reflect the stress concentrations at the tip of the stuck 404 patch at depth but can also enhance them. The geometrical complexities are marked by a 405 wider distribution of the seismicity (Fig. 2d and Supplementary Fig. S4 and S5), 406 associated with a deflection of the San Andreas fault plane (Fig. 8 and Supplementary 407 Fig. S6). Finally, the flower structure shape highlighted in our model (Fig. 6b) is in good 408 agreement with the development of secondary structures at depth that widen towards 409 the surface (Fig. 1d and Supplementary Fig. S6).

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4.1.3. Interpretations and numerical modeling of the twist at Parkfield: 412 strengths and limitations

413 Simpson et al. (2006), using geological and geodetical observations in addition to the 414 2004 aftershock locations of Thurber et al. (2006), interpreted a deflection of the active 415 San Andreas fault near GH-CH along the vertical axis in the upper ~ 6 km from its surface 416 expression as the result of non-elastic behavior of upper crustal rock units. Our 417 observation of a systematic change in fault dip along the entire Parkfield section from 418 MM to GH-CH suggests an additional torque along the horizontal axis that causes the 419 helicoidal nature of the three-dimensional fault plane, as shown in our model (Fig. 6). 420 This shape is not predicted by the 2D numerical model in Simpson et al. (2006), which 421 considers a constant, near-vertical dip at depth to obtain a warp in the fault plane with a major deflection at GH-CH, but not at MM. They suggest that the warp is due to slip in 422 423 the creeping zone (including the Parkfield stuck patch), which results in inelastic deformation that is not released further to the south by large events similar to the 1857
earthquake. Our model of the helicoidal twist is based on the long-term difference in
strength between persistent structures that represent stuck and creeping areas, and
hence does not require inelasticity with such a short time constant.

428 The high friction patch ($\mu \sim 0.6$) is velocity weakening and so exhibits an earthquake 429 cycle. It is embedded in a region of low friction ($\mu \sim 0.2$) which continuously slides in a 430 stable manner. This strength contrast will place a clockwise torque on the system which 431 by slow inelastic deformation over geologic time has twisted the fault. In this regard, our 432 numerical models serve only two purposes: i) to demonstrate that the clockwise torque 433 can arise from the assumed contrast in frictional strength along the Parkfield section 434 and ii) to show that the clockwise torque can be sustained over many earthquake cycles 435 on the high-friction patch.

436 In our model, the shear stress magnitude is significant (~100 MPa) and the 437 correspondend principle between the linear elastic and the linear viscoelastic strain 438 implies, over the long term, that our model's strain field could induce the development 439 of permanent inelastic deformation in the upper part of the crust. This deformation 440 might correspond to the secondary faults and fractures observed at surface and at depth 441 at the tips of the stuck patch. More sophisticated fault and material models would be 442 needed to investigate a causal relationship between the helicoidal twist of the stuck 443 patch and the formation of secondary structures.

444 Our model indicates lateral deformation of the San Andreas fault of 1.5 m to 2 m at MM 445 and 1.3 m to 2.5 m at GH (at surface and at depth, respectively) after 160 m of 446 cumulative slip. Considering the current total cumulative slip around Parkfield since the 447 fault initiation (i.e. ~315 km; (Crowell, 1979; Revenaugh and Reasoner, 1997 and 448 references therein), it would imply a total northeastern fault deformation of ~3 to ~4

km and ~ 2.5 to ~ 5 km at MM and GH, respectively. While this deformation is 449 450 comparable observations at GH (\sim 2 km wide step-over at the surface), it is not the case 451 at MM. However, we note that our model has exaggerated contrast in friction strength 452 between the patch and the creeping section. Increasing the friction coefficient of the 453 creeping section to 0.2, compared to 0.1 used here, would possibly subdue the amount of 454 the northeastern fault deformation. In addition, our model does not include the 455 development of new secondary faults and fractures, which may have been created in the 456 surrounding medium to accommodate the deflection of the San Andreas fault, just like 457 the \sim 5 km wide zone of deformation identified at MM (see Fig. 1 and Thayer, 2006). 458 Additional minor faults might be also present but they have been possibly eroded or 459 hidden under sedimentary deposits.

460 Finally, we remind that our model is a simplified view that cannot exactly reproduce the 461 deformation pattern of the San Andreas fault at Parkfield. The results are sensitive to the 462 initial conditions (i.e. relative location of the stuck patch at depth and connection with 463 the locked section in the south). For instance, we assume a vertical initial fault dip and 464 the model does not reproduce the NE dip of the fault in the creeping section (Fig. 2c). On 465 the other hand, we are not sure that this dip is related to the torque of the fault plane 466 due to the presence of the stuck patch or rather to the initial fault dip of the San Andreas 467 fault. All these parameters were not explored in this study and need to be tested in 468 future models. Our model is used here to show that the right lateral motion and the 469 change in fault sip behavior due to the presence of the stuck patch are in good 470 agreement with the torque highlighted in our observations. The resulting stress and displacement values should be used with caution. 471

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473

4.2. Impact of fault structures on the seismicity behavior and the earthquake cycle

474 Our results confirm that major structural discontinuities, such as secondary reverse and 475 normal faults, are present at the surface near the transitions between the Parkfield and 476 creeping sections at MM, and the Parkfield and Cholame sections near GH-CH (Fig.1). 477 Fine-scale analysis of the seismicity indicates that these structural complexities extend 478 at depth and correlate with the vertices of the elliptical stuck patch that ruptures in M6 479 events (Fig. 2 and 3). It is recognized that aftershocks occur mostly at the edges of 480 regions experiencing high coseismic slip or in areas of low coseismic slip (e.g., Mendoza and Hartzell, 1988), regardless of initial stress conditions. Along the Parkfield section, 481 482 earthquake locations in both the post-seismic and interseismic periods over one and a 483 half seismic cycles delineate the margins of the stuck patch (i.e. transition between 484 locked and creeping areas on the fault; Fig. 2 and 3). Especially, figure 3d updates the 485 common view of the "stationary" seismicity around the stuck patch, going through a new 486 interseismic period, during the decade following the 2004 earthquake. Therefore, our 487 observations indicate that the stress concentrations that surround the stuck patch and 488 therefore the stuck patch itself are long-term permanent features controlling the 489 rupture process.

490 Based on the results from our model (Fig. 7), we schematize in Fig. 9 the state of stress 491 and seismicity distribution on the fault plane. The seismicity distribution reflects 492 significant stress concentrations between creeping regions, which are velocity 493 strengthening and hence frictionally stable, and the locked region (i.e. stuck patch), 494 which is velocity weakening and hence unstable. During interseismic periods (stage 1 in 495 Fig. 9), the stress increases on the stuck patch relative to the surrounding creeping 496 regions and higher stress concentrations are localized at the edge of the stuck patch. The 497 resulting stress concentrations lead to higher seismic activity at the transition between 498 locked and creeping regions. Coseismic slip produces surrounding stress concentrations

during the postseismic period (stage 2 in Fig. 9) which also induces enhanced seismicity
in that area. Then, the stress accumulates again along the locked section during the
following interseismic period (stage 3 in Fig. 9), producing a similar behavior as in stage
1.

503 Our model shows that the long-term deformation of the fault plane is prominent at the 504 along-strike extremities of the stuck patch (Fig. 5 and 6), leading to a concentration of 505 stress and strain that can affect the seismic cycle. These zones also correspond to the 506 nucleation locations of the 1934, 1966, and 2004 earthquakes (Bakun et al., 2005; Bakun 507 and McEvilly, 1979). This behavior is also expected from results of dynamic modeling 508 (Das and Kostrov, 1985) of rupture of an elliptical asperity, where the stress 509 concentrations reach a maximum at the vertices of the semi-major axes of the ellipse. The rupture terminates shortly after it impinges on the velocity-strengthening region. 510 511 This constitutes a stability barrier, as demonstrated by Kaneko et al. (2010). However, 512 recent studies (Noda and Lapusta, 2013) show that some earthquake ruptures might 513 also propagate through creeping areas that were considered to be barriers. The 514 similarity of slip distribution of Parkfield earthquakes seems to suggest that the 515 creeping zones stop rupture propagation at least for M6 and smaller earthquakes. The 516 observation of logarithmically decaying afterslip sequences in the creeping section 517 above the stuck patch following both the 1966 and 2004 earthquakes (e.g. Barbot et al., 518 2009; Langbein et al., 2006; Murray and Langbein, 2006; Scholz et al., 1969) is 519 consistent with the model of Marone et al. (1991), in which a velocity strengthening 520 region overlies a velocity weakening one.

521

522 **5. Conclusions**

523 We present a model based on geological surface observations and seismological 524 observations at depth in which the three-dimensional structural complexities near MM 525 and GH of the San Andreas Fault originate from the presence of a stuck patch 526 surrounded by a creeping zone at Parkfield. The progressive helicoidal twist of the fault 527 plane is due to a torque arising from the contrast in frictional strength between the 528 stuck patch and the surrounding weak creeping area. The complex structures at each 529 end of the helicoid result from the stress concentrations at these locations. They are 530 characterized by complex reverse and normal faulting at the surface and diffuse 531 seismicity at depth. These structural complexities are permanent features, playing an 532 important role in stress concentrations and occurrence of M6 earthquakes at Parkfield.

Faults are laterally segmented along their traces. In the early stages of their evolution, 533 534 the fault segments are separated by geometrical discontinuities possibly formed by 535 small secondary faults and fractures As the fault grows and lengthens through time, fault 536 segments tend to become more and more connected and the lateral discontinuities in between smoothed out (e.g. Manighetti et al., 2015 and references therein). The main 537 538 strand of the San Andreas fault at Parkfield is extraordinarily smooth, as a result of its 539 large cumulative slip (Wesnousky, 1988); the 2-km fault offset at GH-CH being a notable 540 exception. We find that this offset is a constructive edifice resulting in the strength 541 contrast in the fault at this locality. This example of fault roughening with slip is in 542 contrast to the generally assumed smoothing of fully locked faults at they become more 543 mature.

544

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729 FIGURES CAPTIONS

730 Figure 1: a) Map view of central California. White triangles are the 212 stations that 731 recorded the ~17,000 events; b) Fault map along the Parkfield section and relocated 732 seismicity (red dots) from 1966 to 2015. Black circles are events with $M \ge 4$. White stars 733 are the 1966 and 2004 epicenters. Fault traces are shown on SRTM3 topography 734 illuminated from the southwest. Blue lines are quaternary fault traces from the (U.S. 735 Geological Survey and California Geological Survey, 2006); black lines are additional 736 long-term fault traces from this study and Thayer (2006) (dotted black lines when 737 uncertain). SAF: San Andreas Fault; SJF: San Juan Fault; MM: Middle Mountain; GH: Gold 738 Hill; CH: Cholame; c) Zoom into Middle Mountain area. Focal mechanisms (Hardebeck, 739 2010) showing intermediate to low-angle dipping fault motions (< 70°). d) Cross section 740 (AA') showing the relocated seismicity from 1966 to 2015 (red dots) at Middle 741 Mountain. Black crosses are fault traces at surface. The white star represents the 1966 742 hypocenter. Black arrows point possible secondary active strands at depth. Numbered 743 focal mechanisms in c) and d) point out specific thrust (red) and normal (blue) faulting 744 earthquakes (front projection).

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Figure 2: a) Rotated map view of all surface fault traces from fig.1 (black lines) and the relocated seismicity ($M \ge 1$). Blue and red dots are early aftershocks (2 months) of the

748 1966 and 2004 ~M6 earthquakes, respectively. Gray dots are background seismicity 749 (1984 to 2015; see Supplementary Fig. S1 for full catalog). Larger dots are M \geq 4 events 750 since 1979. White stars represent the 1966 and 2004 epicenters. MM: Middle Mountain; 751 GH: Gold Hill. b) Along-strike cross section. Same legend as in a. Circles indicate size of a 752 circular source model with a 3 MPa stress drop. Dashed dark lines are contours of the 753 interseismic slip rate in mm/yr predicted by geodetic inversion (Harris and Segall, 754 1987). c) Variation of the fault strike and dip along the Parkfield section. The mean 755 reference fault strike and dip are 139° and 90° (vertical), respectively. Solid black curve 756 represents the main San Andreas fault strike measured at surface (secondary faults not 757 included). Green and orange curves are inferred strike and dip at depth from the best 758 fitting planes for the 1984-2004 and 2004-2015 interseismic periods, respectively. Gray zones are location of major complexities observed at surface near Middle Mountain and 759 760 Gold Hill-Cholame. d) Half standard deviation from the best fitting planes calculated in 761 each box (see also Supplementary Fig. S4 and S5) for the 1984-2004 and 2004-2015 762 interseismic periods (green and orange shaded curves, respectively).

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Figure 3: Along-strike cross-sections showing projected earthquake density (contours, counts per km²) and cumulative seismic moment (color scale) released along the Parkfield section for a) 1966 and b) 2004 aftershocks, and c) 1984-2004 and d) 2005-2015 periods. White stars are 1966 and 2004 ~M6 hypocenters. Grey dashed lines are coseismic slip distributions (in m) of the a) 1966 (Segall and Du, 1993) and b) 2004 (Bruhat et al., 2011) earthquakes. Gray zones are location of major complexities observed at surface near Middle Mountain and Gold Hill-Cholame.

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Figure 4: (a) Geometry of the model domain and the fault plane. (b) Boundary conditions on the sides and the bottom as well as frictional properties on the fault plane. The top surface is a free surface. The fault section of the model includes creeping sections (in grey) and fully locked zone (in white; stuck patch near Parkfield and Southern Cholame/Fort Tejon section).

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778 Figure 5: (a) Perpendicular displacement of the fault plane toward the northeast and 779 slightly toward the southwest (top) and cumulative fault slip distribution (bottom) 780 around the stuck patch and the high-friction section outlined by the solid black 781 rectangles after 112 m of boundary displacement. (b) Same as in (a) after 160 m of 782 boundary displacement. (c) and (d) Northeast-looking view of the fault plane after 112m and 160 m of boundary displacement, respectively (mesh with the slip magnitudes 783 784 colored in gray scale), and the central plane deformed by the northeast and southwest 785 directed displacements. The solid back rectangles outline the stuck patch and the high-786 friction section.

787

788 Figure 6: (a) Locations of the two vertical cross-sections corresponding to Middle 789 Mountain (MM) and Gold Hill (GH) and one horizontal section going through the center 790 of the stuck patch at the depth of 7.5 km. (b) In-plane displacement magnitudes on the 791 MM cross-section that is displaced by the 500 times amplified in-plane displacement 792 vectors. Black squares are the original locations of the nodes on the central plane and 793 gray squares are the displaced locations of the same nodes after 60 m of boundary 794 displacement. Gray lines show the computational mesh. Thin black lines outline the 795 original rectangular shape of the cross-section. (c) Same as (b) but for the GH cross-

section. (d) Northeast displacement magnitudes plotted on the horizontal sectiondisplaced by the northeast displacement vectors. Symbols are the same as in (b).

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Figure 7: Along-strike shear stress map on the fault plane at two stages of the simulation (after 112 and 160 m of displacement). Note the *x* axis is into the plane and the *y* axis is pointing to the left. So, the right-lateral motion generates negative values of shear stress (σ_{xy}).

803

804 **Figure 8:** Schematic map view (a) and cross section (b) showing the clockwise rotation 805 of the fault plane due to the presence of a stuck patch surrounded by a creeping zone. In 806 (a) the size of the shear couple arrows scales with the amount of shear stress. Positive 807 and minus signs are compressional and extensional regimes, respectively; in (b) the 808 increasing size of dextral motion indicates increasing shearing stresses with depth and 809 thus the clockwise torque (red arrows); c) 3D sketch synthesizing the clockwise twist 810 into a helicoidal fault plane at Parkfield and its relation with changes in fault-slip 811 behavior, structural discontinuities and earthquake distribution; MM: Middle Mountain; 812 GH: Gold Hill; CH: Cholame.

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Figure 9: Sketch showing a simplified view of the Parkfield fault section and along dip distribution of stress and seismicity rates (profile AA') as predicted from a locked patch at mid-crustal depth imbedded in an otherwise creeping fault. Seismic activity is driven by the difference in the stress rate at the edge of the locked patch at every stage of the earthquake cycle.





Perrin et al.; Figure 2





Perrin et al.; Figure 4

(a) Along strike view after 112 m of strike-slip motion (b) After 160 m of strike-slip motion NW SE NW



(c) 3D Northeast-looking view after 112 m of strike-slip motion (d) 3D Northeast-looking view after 160 m of strike-slip motion







(d) Horizontal section at 7.5 km depth



Perrin et al.; Figure 6



Nodes at the original location Nodes displaced by northeastern displacement



(a) After 112 m of strike-slip motion



Max. –120

(b) After 160 m





Perrin et al.; Figure 8

~15 km

Depth



Perrin et al.; Figure 9