

1 **Creep on seismogenic faults: Insights from analogue earthquake**  
2 **experiments**

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18 **Highlights**

- 19 • Stick-slip experiments mimic seismogenic fault behavior  
20 • Creep and earthquakes are not mutually exclusive fault styles  
21 • Interseismic creep varies systematically with fault properties and stress state

22

23

24 **Keywords**

25 *friction, faults, granular materials, aseismic creep, earthquakes, precursors*

26

27 **Abstract**

28 *Tectonic faults display a range of slip behaviors including continuous and episodic*  
29 *slip covering rates of more than 10 orders of magnitude (<mm/a to >m/s). The*  
30 *physical control of such kinematic observations remains ambiguous. To gain insight*  
31 *into the slip behavior of brittle faults we performed laboratory stick-slip experiments*  
32 *using a rock analogue, granular material. We realized conditions under which our*  
33 *seismogenic fault analogue shows a variety of slip behaviors ranging from slow,*  
34 *quasi continuous creep to episodic slow slip to dynamic rupture controlled by a*  
35 *limited number of parameters. We explore a wide parameter space by varying loading*  
36 *rate from those corresponding to interseismic to postseismic rates and normal loads*  
37 *equivalent to hydrostatic to lithostatic conditions at seismogenic depth. The*  
38 *experiments demonstrate that significant interseismic creep and earthquakes may not*  
39 *be mutually exclusive phenomena and that creep signals vary systematically with the*  
40 *fault's seismic potential. Accordingly, the transience of interseismic creep scales with*  
41 *fault strength and seismic coupling as well as with the maturity of the seismic cycle.*  
42 *Loading rate independence of creep signals suggests that mechanical properties of*  
43 *faults (e.g. seismic coupling) can be inferred from shortterm observations (e.g.*  
44 *aftershock sequences). Moreover, we observe the number and size of small episodic*  
45 *slip events to systematically increase towards the end of the seismic cycle providing*  
46 *an observable proxy of the relative shear stress state on seismogenic faults. Modelling*  
47 *the data suggest that for very weak faults in a late stage of their seismic cycle, the*  
48 *observed creep systematics may lead to the chimera of a perennially creeping fault*  
49 *releasing stress by continuous creep and/or transient slow slip instead of large*  
50 *earthquakes.*

51

## 52        **1. Introduction**

53        Faults in the brittle part of the lithosphere may slip at rates ranging from slow,  
54        aseismic ( $< 1 \text{ mm/a}$ ) to fast, seismic ( $> 1\text{m/s}$ ) (Peng and Gomberg, 2010, and  
55        references therein). Moreover they might do so in either continuous (i.e. at constant  
56        rate) or transient fashion (at changing rate). Modern geodetic methods allow  
57        monitoring fault slip rates over time scales long enough to cover a significant part of  
58        the loading history (generally decades) for some fast loading settings like plate  
59        boundaries thereby constraining their kinematic behavior with unprecedented  
60        resolution (Moreno et al., 2010; Shirzaei and Bürgmann, 2013). Accordingly, a suite  
61        of slip behaviors has been observed ranging from continuous creep (e.g., Bokelman  
62        and Kovach, 2003) to transient creep (e.g. precursory and afterslip) (e.g. Bedford et  
63        al., 2013, Schurr et al., 2014) to episodic slip events at various rates (earthquakes,  
64        slow slip and non-volcanic tremor, low frequency earthquakes, creep events) (e.g.  
65        Rogers and Dragert, 2003; Ide et al., 2007). High fluid pressure has been identified as  
66        a controlling factor for slow slip phenomena (e.g., Peng and Gomberg, 2010, Moreno  
67        et al, 2014) but the underlying mechanisms and mechanics controlling which slip  
68        behavior prevails remain under determined. Importantly the physics of such faulting is  
69        often intrinsically undeterminable in nature because of the inaccessibility of the  
70        source and the ambiguity of the geophysical and kinematic observation which can be  
71        fitted by more than one theoretical models and/or set of model parameters.

72        Seismic and aseismic slip behavior are conventionally viewed as mutually exclusive at a  
73        given location through time. Typically “ambivalent” fault slip behaviors are modelled  
74        as a result of the interaction of spatially separated sources, e.g. a seismogenic patch  
75        (asperity) embedded in an aseismic area (barriers) (e.g., Wei et al., 2013). However, a  
76        more integrative view of slow and fast slip phenomena might be possible where the

77 slip behavior is non-unique (e.g. Peng and Gomberg, 2010). Indeed, there is recent  
78 evidence from longterm geodetic observations as well as contrasting geodetic-  
79 seismological versus palaeoseismological observations that given fault areas might be  
80 more variable in their slip behaviors than conventionally believed. In particular we  
81 now have to acknowledge that a particular fault area may show aseismic creep or slow  
82 slip at one time while failing catastrophically in dynamic earthquake ruptures at  
83 others. Examples of spatially overlapping seismic and aseismic fault areas have been  
84 found along the Hayward fault in California U.S. (Lienkaemper et al., 2012, Shirzaei  
85 and Bürgmann, 2013) as well along the subduction megathrusts off Japan (Loveless  
86 and Meade, 2011, Kato et al, 2012) and Chile (Moreno et al, 2010, Ruiz et al, 2014).  
87 As a reaction to such evidence for non-unique slip behavior, existing friction laws  
88 have been adapted for example by allowing aseismic creep at low slip rates but  
89 dynamic weakening at high slip rates, e.g. in the presence of fluids (e.g. Noda and  
90 Lapusta, 2013).

91 We here contribute to the discussion of creep signals by means of experimental  
92 modeling seismogenic fault slip behavior using a lab-scale fault analogue under  
93 conditions relevant to natural faulting. We show that few parameters can control the  
94 rate and stability of fault slip and demonstrate that creeping faults can generate  
95 earthquakes. Showing the systematics by which this happens allows inferring  
96 information on the mechanical properties and state of the fault from kinematic  
97 observations.

## 98 **2. Friction regimes**

99 The most established view on the mechanics of faulting in the brittle regime ( $< c$ .  
100  $350^{\circ}\text{C}$ ) is represented by the rate-and-state dependent friction law (e.g. Scholz, 1998).  
101 This law opens avenues to explain fault slip behavior over a range of rates. In

102 particular, it relates aseismic and seismic fault behavior to an intrinsic velocity-  
103 strengthening and velocity-weakening fault property, respectively. Accordingly, once  
104 static friction is overcome a velocity-weakening fault may weaken dynamically as slip  
105 accelerates resulting in a runaway effect or instability and nucleating an earthquake.  
106 In contrast, an increase of dynamic friction along a velocity strengthening fault  
107 inhibits earthquake nucleation at all times. Importantly, a third regime exists, in which  
108 most of the natural faults might actually be, which is characterized by velocity  
109 weakening under sufficiently low effective normal stress  $\sigma_n'$  (e.g. near the surface or  
110 at high pore fluid pressures). In this regime, which is called the conditionally stable  
111 regime, fault slip is slow and stable under quasi-static loading while it can become  
112 unstable under dynamic loading (acceleration). “Sufficiently” low effective normal  
113 stress in the context of conditional stability means that the externally applied normal  
114 load minus the local pore fluid pressure is below a critical value  $\sigma_c$ :

$$115 \quad \sigma_n' < \sigma_c = kL / -(a-b) \quad (i)$$

116 where  $k$  is the spring stiffness in the original theoretical spring slider framework (or  
117 the stiffness of the medium in which the fault is embedded),  $a$  the instantaneous  
118 change of friction following a loading rate change (so-called direct velocity effect)  
119 and  $b$  the new steady state friction (so-called evolutionary effect) after the loading rate  
120 change which evolves over the characteristic slip distance  $L$  (a physical interpretation  
121 is the size of asperities). The combined parameter  $a-b$  is negative for velocity  
122 weakening interfaces and positive for velocity-strengthening interfaces. Its absolute  
123 values are typically measured in the lab to be in the order of few percent for rocks and  
124 other materials (Scholz, 1998; and references therein).

125

### 126        **3. Analogue earthquake experimental setup**

127        The laboratory-scale analogue earthquake experiments presented here have been  
128        performed in a ring shear tester setup (RST, Figure 1) where a granular material (dry  
129        rice) is sheared rotary in a velocity stepping test under imposed normal loads while  
130        shear stress is measured continuously (e.g. Rosenau et al., 2017, Rudolf et al., 2019).  
131        The rate of laboratory fault slip has been inferred from displacement records derived  
132        by particle image velocimetry (PIV, LaVision Strainmaster ®). For PIV analysis, a 12  
133        bit monochrome charged-coupled device (CCD) camera shot sequential images of the  
134        analogue fault through a transparent shear cell at a frequency of 10 Hz. The particle  
135        motions between successive images are then determined by cross-correlation of  
136        textural differences (i.e., gray values) formed by groups of particles within  
137        interrogation windows using a Fast Fourier Transform algorithm (Adam et al. 2003).  
138        Precision and accuracy of the PIV method is better than 0.1 px of the original image  
139        which scales to the order of micrometer in the presented setup.

140        The stiffness of the loading system ( $\sim 1.3$  kN/mm) together with  $a$ - $b$  ( $\sim -0.015$ ) and  $L$   
141        ( $\sim 2$   $\mu\text{m}$ ) for dry rice (Rosenau et al., 2009) predicts a critical (effective) normal stress  
142        of  $\sigma_c = 8$  kPa. Accordingly, we performed the tests at 1 – 16 kPa normal load to  
143        explore the slip behavior of natural faults across the bifurcation. We refer to the high  
144        (8, 16 kPa) and low (1, 2, 4 kPa) normal stress experiments as strong and weak faults,  
145        respectively.

146        Similarity of the experimental simulation with its natural prototype is ensured by  
147        keeping the following dimensionless numbers the same: (1) the friction coefficient  
148        (ratio between yield strength and normal stress)  $\mu \sim 0.7$ , (Byerlee, 1978) and (2) a  
149        friction rate parameter  $a$ - $b \sim -0.015$  similar to rocks (e.g., Scholz, 1998) as well as (3)

150 a dimensionless stress drop (ratio between rupture slip and length) of  $\Delta\tau^* \sim 10^{-5} - 10^{-4}$   
151 similar to earthquakes (e.g., Scholz, 1989).

152 Applying a stress scale of 1:10.000, the setup generates slip instabilities (aka  
153 “analogue earthquakes”, Figure 2) with stress drops which scale to 1 – 100 MPa in  
154 nature typical of large intra- and interplate earthquakes (Scholz, 1989; Hardebeck and  
155 Aron, 2009) including precursory events of different scale (Figure 3). The strength of  
156 the laboratory fault analogues can be interpreted in two way: Either representing (A)  
157 different crustal depths at a given pore fluid pressure (i.e. weak = shallow, strong =  
158 deep) or (B) representing different pore fluid pressures at a given depth. For example,  
159 at typical seismogenic crustal depths of 5 – 15 km and typical rock densities of 2300 –  
160 2700 kg/m<sup>3</sup>, the experimental normal stresses (10 – 160 MPa) would correspond to  
161 pore fluid pressures of 38 – 96 % lithostatic pressure, i.e. from hydrostatic to near  
162 lithostatic. Time is not explicitly scaled in the experiments but imposed loading rates  
163 cover more than two orders of magnitude (0.1 – 25 mm/min) similar to post- and  
164 interseismic deformation rates in nature (mm/day – mm/year) in order to test possible  
165 time scale dependencies (or independencies) of creep signals.

#### 166 **4. Experimental observations and analysis**

167 Analogue fault slip in our experiments is characterized by quasi-periodic stress drops  
168 (Figure 2). Quasi-periodic stress drops are preceded by smaller, episodic events  
169 (Figure 3). The sizes and recurrence intervals of periodic stress drops are  
170 systematically related to the applied normal load and loading rate (Figure 4). This  
171 observation is consistent with normal load and loading rate both determining the yield  
172 strength according to rate-and-state friction theory (Scholz, 1998). A regular stick-slip  
173 behavior is consistent with a characteristic earthquake model where episodic slip  
174 occurs at a certain stress level determined by the yield strength and causes relaxation



175 to a certain lower stress level determined by the residual friction and the stiffness of  
176 the loading system.

177 Beside periodic and episodic stress drops, representing slip during earthquakes and  
178 slow slip events, a significant amount of long-term laboratory fault slip occurs as  
179 transient creep (accelerating stable slip) between episodic failures. This stable slip  
180 during the “stick”-phase causes the stress curves in Figures 2 and 3 to deviate from a  
181 linear, elastic loading path. Instead of an ideal “saw tooth” pattern characterizing  
182 stress histories of perfect stick-slip, a “shark fin” pattern emerges for the observed  
183 stick-creep-slip. In the experiments, up to 80 % of long-term fault slip might be taken  
184 up by creep at low effective normal stresses resulting in seismic coupling coefficients  
185 (the ratio of seismic to total fault slip) of  $<0.2$  for very weak faults (Figure 2C). At  
186 high normal stresses, seismic coupling increases to  $>0.8$  for strong faults in the  
187 experiments.

188 Detailed inspection of the stress loading paths (Figure 5 A) and interseismic creep  
189 signals (Figure 5 B) and their time-derivates (i.e. loading and slip rates, Figure 5 C  
190 and D) sheds light on the time and stress dependencies of laboratory fault creep.  
191 Accordingly, stress in the inter-event time (which is normalized to a unit interval  
192 here) accumulates in a more transient, non-linear fashion for weak faults than it does  
193 for strong faults (red versus blue curves in Figure 5 A and C). Strong faults show a  
194 stressing rate which is almost consistent with elastic loading except prior to an event  
195 (i.e. runs parallel long-term rate in Figure 5 C) while stressing rates of weak faults  
196 vary by more than an order of magnitude. Slip varies consistently with loading.  
197 Accordingly, slip accumulates in a more non-linear for strong faults than it does for  
198 weak faults (Figure 5 B) covering two orders of magnitude in slip rate versus less than  
199 one, respectively (Figure 5 D).

200 Connecting stress and strain allow us to describe the creep behavior of our fault  
201 analogues as follows: Creep along strong laboratory faults accelerates at rather  
202 constant stressing rate late in the interseismic period leading to episodic failure  
203 (“precursory slip”). Weak faults instead creep at higher rates throughout the  
204 interseismic period but more continuously and at progressively decreasing stressing  
205 rate. Moreover, strong faults reach only about half of the long-term fault slip rate  
206 towards the very end of the loading cycle, whereas weak faults may creep at almost  
207 the long-term rate for the second half of the loading cycle.

208 In order to analyze the creep behavior systematically as controlled by extrinsic factors  
209 (normal stress and loading rate) we attempted to quantify the non-linearity (or  
210 transience) of stress and slip accumulation by a single, dimensionless parameter.  
211 Therefore we calculated the area beneath the normalized stress and strain  
212 accumulation curves in Figure 5 A and B, respectively, which we call the unit stress  
213 and unit strain integrals (Figure 5E). Clearly, these measures of transience decrease  
214 systematically with increasing applied normal stress or fault strength as expected from  
215 the observations before. However, they do not correlate with loading rate, an  
216 observation that is not intuitive but useful as will be discussed below. The positive  
217 correlation between the unit stress and slip integrals (Figure 5F) indicates the  
218 consistency of our independent stress and strain observations and is a direct result of  
219 the intrinsic velocity weakening behavior of the laboratory fault.

220 Irrespective of fault strength, episodic slip events of various speeds occur at high  
221 stress level modulating the interseismic creep signal in the late stage of the analogue  
222 seismic cycle (Figure 3). Preliminary analysis suggests that these precursor events  
223 increase systematically in number and size as the fault evolves towards failure.

224

## 225        **5. Discussion**

### 226                **5.1 Inversion of fault properties and state from creep signals**

227    The observation of continuous and transient creep signals as well as episodic slow  
228    slips which are systematically linked to fault properties and maturity of the loading  
229    cycle or stress level but independent of loading rate bear important implications for  
230    the interpretation of fault creep records as observable proxies for fault strength and  
231    seismic potential. Fault creep records in nature are generally short with respect to the  
232    seismic cycle. The results obtained here suggest that any creep record, though only a  
233    snapshot of the full seismic cycle, might bear important information on long-term  
234    fault properties and hazardous behavior.

235    Using the analog fault observations from the here presented experiments, an empirical  
236    inversion scheme as proposed in Figure 6 can be applied, where inaccessible fault  
237    properties like fault strength, seismic coupling, stress drop and recurrence interval can  
238    be inferred from the observable transience of interseismic creep signals. Here, creep  
239    transience (CT) is defined as

$$240 \qquad \qquad \qquad CT = 2 \cdot (1 - 2 \cdot \text{unit slip integral}) \qquad \qquad \qquad \text{(ii)}$$

241    in order to derive a dimensionless (and therefore scale-independent) parameter which  
242    varies between 0 (linear strain accumulation) and 1 (non-linear, highly transient strain  
243    accumulation).

244    Linear regression analysis of the experimentally derived data plotted in such a scheme  
245    indicates a significant correlation between creep signals and fault properties and  
246    behavior but independence of loading rate. More specifically, fault strength, seismic  
247    coupling, stress drop as well as recurrence period show a positive linear or log-linear  
248    dependency with CT ( $R^2 > 0.6 - 0.8$ ).

249 Importantly, no significant correlations exist between any of the parameters with  
250 loading rate. This is indicated by the rather horizontal or scattered distribution of data  
251 from subsets with the same fault strength measured at different velocities in Figure 6  
252 as well as the collapse of time-series data from such subsets in Figure 5. The fact that  
253 the systematics found experimentally are loading rate independent suggest that short-  
254 term observations can be extrapolated to larger earthquakes and longer recurrence  
255 intervals. I.e. this timescale independency opens the opportunity to generalize fault  
256 properties or behavior derived during aftershocks sequences or earthquake swarms or  
257 from repeating events to longterm (multiple seismic cycles) fault behavior.

258 An observation not quantified in detail here is the occurrence of precursor slip events  
259 of different scale and velocity which systematically increase in number and size  
260 towards the end of a seismic cycle (Figure 3). Several large earthquakes in subduction  
261 zones have actually been preceded by accelerating foreshock activity (e.g. Bouchon et  
262 al., 2013). Especially the recent 2014 8.1 Pisagua earthquake offshore Chile showed  
263 accelerating foreshock activity with a decrease in b-value (representing an increase in  
264 the number of large events relative to small events) over the decade preceding the  
265 main shock (Schurr et al., 2014). If such a systematic behavior can be generalized and  
266 physically explained it should lead to a better ability to forecast earthquakes.

## 267 **5.2 Revisiting creep records along the San Andreas Fault**

268 In order to test and apply our proposed inversion scheme, we use the longest creep  
269 records available and revisit the San Andreas Fault data. California creepmeters have  
270 been installed across the San Andreas Fault in the late 1960s (Schulz et al., 1982),  
271 geodetic surveys took place since the mid-1970s (Burford and Harsh, 1980; Lisowski  
272 and Prescott, 1981) and surface velocities from space-geodetic measurements are  
273 available since about a decade (e.g., Bürgmann et al., 2000; Titus et al., 2006). For a

274 mean recurrence interval of large Californian earthquakes of about  $150 \pm 50$  years  
275 along any SAF segment (e.g. Zielke et al., 2010), the observation time frame  
276 generally represents less than half of the seismic cycle length. Nevertheless, the  
277 records are probably the best data we can get today.

278 Seismic and aseismic strike-slip along the central SAF (cSAF) accounts for most of  
279 the Pacific-Great Valley microplate relative motion in central California (Thatcher,  
280 1979; Lisowski and Prescott, 1981, Titus et al., 2006; Rolandone et al., 2008; Ryder  
281 and Bürgmann, 2008). As suggested by over 40 years of creep and earthquake  
282 records, the central section of the cSAF creeps continuously at a decadal scale at  
283 about 28 mm/a at seismogenic depth (0 – 12 km, Schulz et al., 1982, Titus et al.,  
284 2006, Rolandone et al., 2008). This long-term creep is modulated by shorter term  
285 transients presumably very shallow (< 5 km) and related to earthquakes (Lisowski and  
286 Prescott, 1981; Thurber, 1996). At seismogenic depths repeating microearthquakes  
287 occur (Nadeau and McEvilly, 2004) indicating that locally and/or transiently, velocity  
288 weakening behavior is established along the fault. Noticeably, the current creep of  
289 cSAF is only about 80 – 90 % of the far-field, tectonic loading rate (31 – 35 mm/a,  
290 Titus et al., 2006, Rolandone et al., 2008; Ryder and Bürgmann, 2008) suggesting a  
291 slip deficit of few millimeter accumulating each year. Right-lateral shear strains in the  
292 sidewalls of the cSAF are evidently very small (Rolandone et al., 2008, Savage, 2009)  
293 suggesting a small stressing rate. Episodic slow slip events as they occur late in the  
294 interseismic period in our experiments (Figure 3) have been reported as potential  
295 earthquake pre-cursors along the SAF by Thurber (1996) and Thurber and Sessions  
296 (1998) based on temporal cross-correlation of creepmeter records and seismological  
297 catalogues. Though the correlations they found were statistically significant, the  
298 feedback mechanism remained unclear. Noticeably, they did not find a clear spatial

299 relation between the loci of creep and earthquakes which would be required by our  
300 model. Moreover, they assigned creep to the very shallow crust (<5 km) and not to  
301 seimogenic depths. Whilst the adjoining segments ruptured in large earthquakes in  
302 1906 (San Francisco) and 1857 (Fort Tejon), the creeping section of the cSAF has not  
303 experienced large earthquakes in the historic past (~300 years).

304 In the light of the experiments done in this study the key question is: Does the absence  
305 of large earthquakes, the high and continuous creep rates as well as the low shear  
306 strain accumulation serves as a good indicator that this fault segment poses no seismic  
307 hazard?

308 Applying the empirical inversion scheme established above (Figure 6), we would  
309 infer first that the creeping section of the cSAF is a very weak fault based on the  
310 rather linear slip accumulation signal (Schulz et al., 1982, Titus et al., 2006) and low  
311 stressing rate (Rolandone et al., 2008, Savage, 2009). This is consistent with previous  
312 findings based on the observation of low resolved shear stresses along the creeping  
313 section and absence of a heat flow anomaly (Brune et al., 1969, Lachenbruch and  
314 Sass, 1980, Zoback et al., 1987).

315 The cSAF shows therefore kinematic similarity to our weak fault analogue  
316 characterized dynamically by low seismic coupling and small stress drops during  
317 earthquakes. This may however not mean that the seismic potential is low. In contrast:  
318 Because stress drop is only a weak measure of earthquake size, which scales  
319 dominantly with the rupture area, and because low seismic coupling (or vice versa a  
320 large amount of interseismic creep) just stretches the recurrence intervals of  
321 potentially large earthquakes. We will elaborate on this effect in the next section.

322 **5.3 Modelling the effect of creep on recurrence time and the chimera of**  
323 **perennially creeping faults**

324 Because of the empirically found correlation between fault strength and creep, the net  
325 effect of creep on the recurrence interval of earthquakes should not only take into  
326 account the stretching of the recurrence interval due to creep but also a modification  
327 of recurrence interval due to changes in strengths (Figure 4). Such a scenario is  
328 illustrated in Figure 7.

329 Quantitatively, creep lengthens the (effective) recurrence interval to

330 
$$t^* = 1/(1-\text{creep}). \quad (\text{iii})$$

331 For example a fault where 50 % of longterm slip is accommodated aseismically  
332 requires twice as much time to reach a certain stress level again. However, because  
333 creep correlates with fault weakness and weaker faults fail at lower stress level in  
334 quicker succession for the same far field stressing rate (Figure 4), this lengthening  
335 effect is to some degree counterbalanced by shorter recurrence intervals.

336 In Figure 7 we plot the effective recurrence time observed in our experiments in  
337 relation to creep on faults of variable strength and model the data as the combined  
338 result of the competing effects of “creep lengthening” (according to eq. (iii)) and  
339 “weakness shortening”. The latter effect is taken into account by fitting an  
340 exponential relation of the form

341 
$$t^{**} = e^{(-A \times \text{creep})} \quad (\text{iv})$$

342 to the data. Parameter  $A$  is an empirically derived proxy for the relation between  
343 strength and recurrence interval and varies between 4 and 6 in our example. The net  
344 effect of “creep lengthening” and “weakness shortening” of recurrence intervals, i.e.  
345 the effective recurrence interval, is then simply

346 
$$t = t^* \times t^{**} = 1/(1-\text{creep}) e^{(-A \times \text{creep})}. \quad (\text{v})$$

347 For the parameter space realized in our experiments recurrence time is always shorter  
348 than on faults without creep, i.e. the weakness effect dominates the recurrence  
349 behavior such that more creeping faults have systematically shorter recurrence times.  
350 However, at least theoretically our model predicts for very weak faults (not realized in  
351 our experiments) with very low seismic coupling coefficients and very high creep  
352 amounts, the lengthening effect should start dominating and consequently the  
353 effective recurrence intervals should become longer than without creep. For creep  
354 amounts exceeding 98% effective recurrence times may well exceed any historical  
355 record for fast creeping faults (Figure 7). In the extreme such a seismically nearly  
356 uncoupled, very weak fault appears as seismically silent over many human  
357 generations – obviously a chimera.

#### 358 **5.4 Creep on continental vs. subduction megathrusts**

359 Locking pattern of continental and subduction megathrusts show a striking qualitative  
360 difference: While continental megathrusts, e.g. the Himalayan main thrust, show  
361 homogeneous and high locking with little interseismic creep (Stevens and Avouac,  
362 2015), subduction megathrusts, like the Chilean subduction zone, show a patchy  
363 locking pattern indicating a significant amount of creep (e.g. Saillard et al., 2017).  
364 According to our experiments, and in line with theory, such a qualitative difference  
365 can be explained by higher amounts water entrained into subduction megathrust  
366 compared to continental settings, lowering the effective normal load and this  
367 enhancing creep. However, other explanations exist like differences in lithology and  
368 even lack of offshore geodetic coverage.

369



370 **6 Conclusion**

371 Based on stick-slip experiments using a lab-scale fault analogue, we explored the slip  
372 behavior of seismogenic faults and tested the potential to derive information on fault  
373 properties and state from kinematic observables. We showed that the stress buildup  
374 between episodic failures (analogue earthquakes) is non-linear and anti-correlated  
375 with the creep signals. According to our experiments the transience of stress buildup  
376 and creep is controlled primarily by fault normal stress, i.e. related to frictional  
377 strength and/or pore-fluid pressure, and systematically reflect the seismic coupling  
378 coefficient and maturity of the seismic cycle. Application of these systematics to the  
379 creeping section of the central San Andreas fault suggests that this fault branch may  
380 not be aseismic on the long term (millennia scale) but is in a late stage of a seismic  
381 cycle which exceeds historic records. The qualitative difference in creep on  
382 megathrusts between homogeneously fully locked continental versus heterogeneously  
383 locked subduction megathrusts may be similarly explained by the presence of water in  
384 oceanic settings.

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388 project B01.

389

390 **Figure Captions**

391 **Figure 1:** Analogue earthquake experimental setup: (A) side (camera) view of the  
392 sample (rice) in a transparent shear cell in situ in the ring-shear tester, boundary  
393 conditions and observables indicated; PIV velocities are representative of a slip event.  
394 (B) sketch of the ring-shear tester setup (modified from Schulze (2003)) with PIV  
395 camera position indicated.

396 **Figure 2:** Stress and strain time-series of laboratory faults: (A) Stress time series  
397 measured during velocity stepping tests under variable normal loads simulating  
398 seismic and aseismic slip along very weak to strong fault slip. Note the periodic stress  
399 drops representing analogue earthquakes. (B) Slip time series for very weak and  
400 strong faults derived by PIV. (C) Variation of seismic coupling over the parameter  
401 space tested here. Note the sensitivity of seismic coupling to normal load and  
402 insensitivity to loading rate.

403 **Figure 3:** Examples of precursory slip events along laboratory faults (from Rosenau  
404 et al., 2017): (A) stress time series, (B) Histogram of number of slow slip events per  
405 unit interseismic time interval. Note the increase of precursory events in size and  
406 number towards the end of the seismic cycle.

407 **Figure 4:** Dependency of recurrence interval and stress drop on loading rate and  
408 normal load over the parameters space tested here.

409 **Figure 5:** Systematics of interseismic stress-strain relationships for laboratory faults:  
410 (A) interseismic stress accumulation (normalized), (B) interseismic slip accumulation  
411 (normalized), (C) interseismic stress rate (normalized), (D) interseismic slip rate  
412 (normalized), (E) Variation of unit stress and slip integrals over the parameter space

413 tested here, (F) correlation of unit stress and slip integrals indicating velocity  
414 weakening behaviour.

415 **Figure 6:** Dependency of creep signal transience on laboratory fault properties: (A)  
416 fault strength as a function of creep transience, (B) seismic coupling as a function of  
417 creep transience, (C) stress drop as a function of creep transience, (D) recurrence  
418 period as a function of creep transience.

419 **Figure 7:** Modelling the effect of fault creep and strength on recurrence time of  
420 earthquakes. Experimental data are fitted by theoretical model taking into account two  
421 competing effect: Fault creep lengthens recurrence intervals (“creep lengthening  
422 effect”) while weakening faults should shorten recurrence intervals (“weakness  
423 shortening effect”). The effective recurrence is dominated by the weakness effect for  
424 faults creeping up to 98%. However, faults which accumulate >98 % of fault slip  
425 aseismically may still generate earthquakes with recurrence periods exceeding  
426 historical records (California earthquake history shown as example).

427

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