



**HAL**  
open science

## Fast and Slow Slip Events Emerge Due to Fault Geometrical Complexity

Pierre Romanet, Harsha S. Bhat, Romain Jolivet, Raúl Madariaga

► **To cite this version:**

Pierre Romanet, Harsha S. Bhat, Romain Jolivet, Raúl Madariaga. Fast and Slow Slip Events Emerge Due to Fault Geometrical Complexity. *Geophysical Research Letters*, 2018, 45 (10), pp.4809-4819. 10.1029/2018GL077579 . hal-02185462

**HAL Id: hal-02185462**

**<https://hal.science/hal-02185462>**

Submitted on 26 Aug 2019

**HAL** is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

# Fast and slow slip events emerge due to fault geometrical complexity

Pierre Romanet<sup>1,2\*</sup>, Harsha S. Bhat<sup>2</sup>, Romain Jolivet<sup>2</sup>, Raúl Madariaga<sup>2</sup>

<sup>1</sup>Institut de Physique du Globe de Paris, CNRS-UMR 7154, Sorbonne Paris Cité, Paris 75005, France

<sup>2</sup>Laboratoire de Géologie, École Normale Supérieure, CNRS-UMR 8538, PSL Research University, Paris 75005, France.

## Key Points:

- Fault geometry can be a natural source of slip complexity in earthquake cycle modeling, resulting in slow slip events (SSE) and earthquakes.
- A simple two overlapping fault model produces different observed scaling laws for earthquakes and for slow slip events.
- All observed complexities emerge with uniform loading and rate weakening friction properties on the fault.

This is the accepted version of the manuscript.  
There might be minor differences with the published article  
following editing, proof-reading and publication

---

\*Now at Department of Earth and Planetary Sciences, School of Science, The University of Tokyo

Corresponding author: Pierre Romanet, romanet@geologie.ens.fr

**Abstract**

Active faults release elastic strain energy via a whole continuum of modes of slip, ranging from devastating earthquakes to Slow Slip Events and persistent creep. Understanding the mechanisms controlling the occurrence of rapid, dynamic slip radiating seismic waves (i.e. earthquakes) or slow, silent slip (i.e. SSEs) is a fundamental point in the estimation of seismic hazard along subduction zones. Using the numerical implementation of a simple rate-weakening fault model, we show that the simplest of fault geometrical complexities with uniform rate weakening friction properties give rise to both slow slip events and fast earthquakes without appealing to complex rheologies or mechanisms. We argue that the spontaneous occurrence, the characteristics and the scaling relationship of SSEs and earthquakes emerge from geometrical complexities. The geometry of active faults should be considered as a complementary mechanism to current numerical models of slow slip events and fast earthquakes.

**1 Introduction**

Since their discovery in the late nineties, Slow-Slip Events (SSE) have been widely observed along various subduction zones (Central Ecuador [Vallee *et al.*, 2013], Southwest Japan [Hirose *et al.*, 1999], Guerrero [Lowry *et al.*, 2001], Cascadia [Dragert *et al.*, 2001; Rogers and Dragert, 2003], Hikurangi [Douglas *et al.*, 2005], Northern Chile [Ruiz *et al.*, 2014] and others). The discovery of SSEs mainly came from the development and the installation of networks of permanent GPS stations around subduction zones. Although GPS is still nowadays the main SSE detection tool, new observations now allow for the detection of slow-slip, like InSAR [Rousset *et al.*, 2016; Jolivet *et al.*, 2013], networks of sea-bottom pressure gauge [Ito *et al.*, 2013; Wallace *et al.*, 2016] or, indirectly, via the migration of microseismicity, repeating earthquakes and tremors [Igarashi *et al.*, 2003; Kato *et al.*, 2012], thus increasing significantly the probability of their detection.

SSEs, like earthquakes, correspond to an accelerating slip front propagating along a fault. However, unlike earthquakes, SSEs themselves do not radiate any detectable seismic waves and are hence sometimes nicknamed “silent events”. Until the discovery of SSEs, it was thought that only earthquakes release the accumulated strain energy along a fault. Since SSEs also contribute to this release of energy, they should play an important role in the estimation of seismic hazard along subduction zones [Obara and Kato, 2016]. In addition, SSEs exhibit very specific characteristics. Their propagation speed

45 along the fault (about 0.5 km/h in Cascadia [Dragert *et al.*, 2004] to about 1 km/day in  
46 Mexico [Franco *et al.*, 2005]) contrasts with the rupture propagation speed of earthquakes  
47 (at about 3 km/s). The slip velocity of SSEs (from about 1mm/yr in the Bungo Channel,  
48 Japan to about 1 m/year in Cascadia) is around one or two orders of magnitude greater  
49 than plate convergence rates but orders of magnitude smaller than earthquakes slip rates  
50 (of the order of 1m/s) [Schwartz and Rokosky, 2007].

51 Although the exact influence of SSEs in the seismic cycle is not yet fully under-  
52 stood, they seem closely related to earthquakes. Several seismic and geodetic observa-  
53 tions suggest that SSEs may have happened just before and in regions overlapping with  
54 earthquakes. The 2011  $M_w$  9.0 Tohoku-Oki event and the 2014  $M_w$  8.1 Iquique event  
55 are two examples in subduction zones where a SSE apparently occurred just before the  
56 earthquake, within a region overlapping with the area where seismic slip nucleated [Kato  
57 *et al.*, 2012; Brodsky and Lay, 2014; Ruiz *et al.*, 2014; Mavrommatis *et al.*, 2015]. More  
58 recently, geodetic evidence of a large SSE triggering an earthquake was pointed out in the  
59 Guerrero subduction zone [Radiguet *et al.*, 2016]. There are also suggestions that SSEs  
60 may be triggered by earthquakes either by stress-waves and/or static stress transfer [Itaba  
61 and Ando, 2011; Zigone *et al.*, 2012; Kato *et al.*, 2014; Wallace *et al.*, 2017]. On the other  
62 hand some SSEs occur without an accompanying large earthquake as in the Cascadia sub-  
63 duction zone, where SSEs occur periodically [Rogers and Dragert, 2003], or in the Hiku-  
64 rangi subduction zone [Wallace *et al.*, 2016]. From the above examples, it seems that there  
65 may or may not be a connection between slow slip events and fast earthquakes. Some au-  
66 thors [Obara and Kato, 2016, for e.g.] have suggested that slow slip events, because of  
67 their sensitivity to very small stress perturbations, can act as a stress meter of the current  
68 stress in the crust. However, this still needs to be confirmed. Also, the exact role of SSE's  
69 in hazard assessment remains largely unknown. All SSEs have the same direction of slip  
70 as earthquakes, i.e. opposite to the plate convergence direction, and are accompanied by  
71 a positive stress drop which corresponds to a reduction in the accumulated strain energy.  
72 In the absence of external forcing mechanism, this necessitates SSEs to occur in a slip, or  
73 slip rate, weakening region which is also prone to rupture as a fast dynamic event. These  
74 observations, put together, raise the first question. *What physical mechanism explains slow-*  
75 *slip and fast, dynamic earthquakes occurring under similar frictional boundary conditions*  
76 *along active faults?* Our key finding is that fault geometrical complexity gives rise to the

77 variety of modes of slip along an active fault without any other complex mechanism in-  
78 volved.

79 Furthermore, earthquakes and SSEs seem to follow different scaling laws [*Ide et al.*,  
80 2007], which remain out of reach of numerical models until now [*Ide*, 2014]. The seismic  
81 moment of earthquakes scales with the cube of their duration ( $M \propto T^3$ ) whereas the cor-  
82 responding moment of SSEs is proportional to their duration ( $M \propto T$ ), raising the second  
83 question. *Is such different scaling a general feature of earthquakes and SSEs, highlighting*  
84 *different physical mechanisms* [*Ide et al.*, 2008; *Peng and Gomberg*, 2010; *Ide*, 2014]? We  
85 address the above questions using physics-based numerical modeling of active faults gov-  
86 erned by rate-and-state friction [*Dieterich*, 1978] and develop a unified framework that  
87 addresses all the observations about earthquakes and SSEs mentioned above.

## 88 **2 Modeling slow, aseismic slip**

89 SSEs were discovered to emerge spontaneously from numerical models in the rate-  
90 and-state framework for the modeling of subduction zones [*Liu and Rice*, 2005, 2007].  
91 In this framework, fault areas with weakening properties will preferentially host seismic  
92 slip (i.e. earthquakes) while strengthening regions will host stable continuous creep or  
93 post-seismic slip. Numerical experiments and theoretical works have shown that the main  
94 physical control on the emergence of SSEs in models is how the characteristic length of a  
95 weakening patch [*Ruina*, 1983; *Rice*, 1983; *Dieterich*, 1992; *Rubin and Ampuero*, 2005]  
96 compares to the specific nucleation length scale [*Liu and Rice*, 2005; *Rubin*, 2008]. If  
97 the length of a fault patch is large compared to the nucleation length scale, earthquakes  
98 have enough room to grow and become dynamic, so this fault patch will generate only  
99 dynamic, seismic events. If the length of the fault is small compared to this length scale,  
100 earthquakes can never grow large enough to become dynamic or no events will occur at  
101 all (i.e. permanent creep). It is therefore necessary, under this framework, to tune for the  
102 right fault length compared to the nucleation length scale to allow modeling of both slow  
103 and fast ruptures. Given the observed spatial size over which some SSEs propagate i.e.  
104 on the order of tens of kilometers, this would lead to unrealistically large nucleation sizes,  
105 preventing the occurrence of any earthquakes. A possible explanation for such large nucle-  
106 ation lengths could be the presence of high-pressure pore fluids released during metamor-  
107 phic dehydration reactions. However it has been shown recently that regions of high fluid  
108 pressure and slow slip events do not always overlap along all the subduction zones [*Saffer*

109 *and Wallace, 2015*]. One solution to overcome this issue is to appeal to other compet-  
110 ing frictional mechanisms like dilatant-strengthening [*Segall and Rice, 1995; Rubin, 2008;*  
111 *Segall et al., 2010*] with or without thermal-pressurization [*Segall and Bradley, 2012*]. Al-  
112 though we do not include these additional frictional mechanisms in our modeling below,  
113 we acknowledge that it would broaden the range over which we are able to observe slow-  
114 slip.

115 As the above models suggest, a set of competing mechanisms are required for slow-  
116 slip and earthquakes to coexist. However, there is one ubiquitous feature that is often ig-  
117 nored for computational reasons: the geometric complexity of active faults. Indeed, faults  
118 are rarely planar over length scales of tens of kilometers and in fact, fault segmentation  
119 and geometric complexity are visible at multiple scales [*Candela et al., 2012*]. Subduc-  
120 tion zones also show geometrical complexities like subducting seamounts [*Das and Watts,*  
121 *2009*]. It is also known that subduction zones have large normal faults that connect the  
122 main slab and can sometimes be reactivated during seismic events [*Hicks and Rietbrock,*  
123 *2015; Hubbard et al., 2015*].

124 This non-planarity of faults should introduce a natural stress based interaction be-  
125 tween faults. Several lines of evidence suggest that geometric complexity should be con-  
126 sidered in conjunction with various observed slip dynamics. Aseismic slip has been ob-  
127 served with earthquake swarms in the northern Apennines (Italy) along splay faults [*Gua-*  
128 *landi et al., 2017*]. It has been detected along the Haiyuan fault (China) [*Jolivet et al.,*  
129 *2013*], the North Anatolian Fault [*Rousset et al., 2016; Bilham et al., 2016*] and, in earlier  
130 publications, along the San Andreas Fault [*Murray and Segall, 2005*]. SSE's have been ob-  
131 served in the very shallow part of subduction zones, such as in Hikurangi [*Wallace et al.,*  
132 *2016*] and Nankai [*Araki et al., 2017*], among others. The only known common ingredi-  
133 ent of all of these different seismotectonic settings is the geometrical complexity of faults  
134 across scales.

135 In this work, we have restricted ourself to only one type of geometric complexity  
136 i. e. two overlapping faults. Of course, this geometry cannot be interpreted directly as a  
137 subduction zone or any other natural setting. However, we suggest that if this simple ge-  
138 ometry can give rise to a complex slip behaviour in the seismic cycle then a more realistic  
139 description of fault zones with multiple slip surfaces should not be ignored.

### 140 3 Model set-up

141 Our aim is to test the influence of fault geometry on the behavior of slip along a  
 142 fault. We build a conceptual model in which fault slip is controlled by an unstable fric-  
 143 tional rheology (rate weakening) without any lateral variation. Doing so, we introduce no  
 144 a priori complexity in initial and boundary conditions. We load the faults with constant  
 145 stress loading rate and observe the variety of modes of slip.

146 In our conceptual model, we consider two overlapping faults of the same length  $L$   
 147 (see geometry in Fig. 1). This geometry is chosen to illustrate the effect of complex stress  
 148 interactions between neighboring faults or fault segments and is in no way supposed to  
 149 be interpreted as the only geometrical configuration of faults in a fault network. Friction  
 150 on both faults is controlled by rate-and-state friction with aging state evolution. Frictional  
 151 resistance decreases with increasing slip rate and is spatially uniform, i.e. the fault is rate-  
 152 weakening. Loading is imposed using a constant rate of shear stress increase on the fault.  
 153 We model elastic interactions using out-of-plane static stress interactions with a radia-  
 154 tion damping approximation [Rice, 1993]. The computation of static stress interactions  
 155 is accelerated using the Fast Multipole Method, allowing us to compute all stages of the  
 156 earthquake cycle in a tractable computational time [Greengard and Rokhlin, 1987; Carrier  
 157 *et al.*, 1988] (See Methods section for more details).

158 To better understand the role of multi-fault interactions, we explore the influence of  
 159 the distance between faults,  $D$ , the length of the faults,  $L$ , and the ratio of the constitu-  
 160 tive frictional parameters,  $a/b$ . For rate-weakening faults,  $a/b$  ranges between 0 and 1.  
 161 Because of the importance of the nucleation length scale  $L_{nuc}$  in this problem, all geomet-  
 162 rical parameter are non-dimensionalized by  $L_{nuc}$ ,

$$L_{nuc} = \frac{2}{\pi} \frac{\mu D_c}{\sigma_n b (1 - a/b)^2} \quad ; \quad a/b \rightarrow 1 \quad (1)$$

163 where,  $a$  and  $b$  are rate-and-state constitutive friction parameters,  $D_c$  is the characteristic  
 164 slip distance,  $\mu$  is the shear modulus of the medium and  $\sigma_n$  the normal stress acting on  
 165 the fault [Rubin and Ampuero, 2005; Viesca, 2016]. This formulation provides good in-  
 166 sights on the nucleation phase of earthquakes along a fault that is mildly rate-weakening  
 167 ( $a/b \rightarrow 1$ ).

168 For computational reasons, we restrict our experiments to fault lengths  $L/L_{nuc} \in$   
 169  $\{1, 2, 3, 4\}$ . Our parameter space includes also distances between faults  $D/L_{nuc} \in \{0.1, 0.5, 1, 2, 3, 4\}$ ,

170 and constitutive parameters  $a/b \in \{0.7, 0.8, 0.85, 0.90, 0.95\}$ . For illustrative purposes  
 171 we provide a table of dimensional values of  $L$  and  $D$  in the supplementary section. The  
 172 smallest faults are 200 m long separated by distance of 21 m. The largest faults are about  
 173 20 km long separated by a distance of about 2 km. In fact, it is possible to distinguish be-  
 174 tween different domains of behavior, that mainly depend on  $a/b$ ,  $L/L_{nuc}$  and the scaled  
 175 distance between the faults  $D/L_{nuc}$ .

#### 176 **4 Results**

177 For each of the parameters identified above, we initiate the model, and compute slip  
 178 velocity over time (Fig. 1). We observe cycles of quiescence and earthquakes as expected  
 179 for a rate-weakening rheology but, unlike in a model with a single, flat fault with no ge-  
 180 ometrical complexity, we also observe episodes during which slip is slow. In our con-  
 181 ceptual model, we see regular earthquakes with a clear nucleation, dynamic and afterslip  
 182 phases and these events happen without any evident periodicity. We observe what would  
 183 be considered in nature as the slow nucleation of earthquakes, the slow phase of recov-  
 184 ery following an earthquake, earthquakes of variable slip duration and velocity and slow  
 185 slip events. It appears then, that the sole introduction of a simple geometrical complex-  
 186 ity leads to the emergence of the complete range of modes of slip, even with a uniform  
 187 rate-weakening rheology. Slow-slip events emerge spontaneously without prescribing the  
 188 necessary conditions for slow slip. In our model, a fault that slipped seismically can also  
 189 potentially host slow slip, as in the region of overlap of co- and post-seismic slip or along  
 190 the shallow portion of a creeping fault [Wallace *et al.*, 2016; Rousset *et al.*, 2016]. Once  
 191 again, without the introduction of a second fault, and its associated stress perturbations,  
 192 the fault behaves like a simple spring-slider system with weakening properties, with simi-  
 193 lar earthquakes happening periodically (see Figure S3 in Supp. Mat.).

194 We believe the choice of such geometry brings realistic perturbations in stress along  
 195 the fault and these perturbations lead to the emergence of the observed variety of modes  
 196 of slip. Fig. 1 illustrates the complexity that emerges by only appealing to stress pertur-  
 197 bations from a neighboring fault and/or non planarity of the fault. Now considering that  
 198 faults are geometrically complex at all scales, it appears natural to extend this conclusion  
 199 and consider that the whole range of modes of slip observed in nature may result, among  
 200 other mechanisms, from these geometrically-induced stress complexities. In addition, it  
 201 may be safe to think that models that do not include such complexities will require ad-hoc

202 tuning, which might not be necessary, to reproduce observations. We have not yet identi-  
 203 fied the precise conditions leading to an earthquake or a slow slip event, but clues should  
 204 be found in the analysis of the evolution of stresses and state variable along the fault.

#### 205 **4.1 A phase diagram of slip**

206 We allow our model to undergo multiple earthquake cycles before measuring slip  
 207 and rupture velocity of each slow and dynamic event. We identify SSEs and earthquakes  
 208 based on their slip and rupture velocity. SSEs are events with a slip velocity  $V$  in the  
 209 range of  $1\mu\text{m/s}$  to  $1\text{ mm/s}$  and a rupture velocity  $V_{rup}$  lower than  $0.001c_s$ , where  $c_s$  is the  
 210 shear wave speed. Earthquakes are events with a slip velocity greater than  $1\text{ mm/s}$  and a  
 211 rupture velocity greater than  $0.001c_s$ . We also define nucleation as the moment before an  
 212 earthquake, where slip velocity is higher than  $1\mu\text{m/s}$  until it reached  $1\text{ mm/s}$ . We purpose-  
 213 fully chose a relatively small threshold value for rupture velocity, because quasi-dynamic  
 214 simulations lead to much slower rupture velocity than dynamic simulations [Thomas *et al.*,  
 215 2009]. As our faults are one dimensional, we define the equivalent moment for a seismic  
 216 or aseismic event as  $M = \mu\bar{D}L_{rup} \times 1\text{ km}$ , where  $L_{rup}$  is the total length of the fault that  
 217 slipped during an event (SSE or earthquake) and  $\bar{D}$  is the slip averaged over the length  
 218  $L_{rup}$ . For earthquakes, we compute separately the seismic moment during the nucleation  
 219 phase and the dynamic phase. For SSEs, moment accounts for the entire duration when  
 220 the slip velocity exceeds  $1\mu\text{m/s}$ . We obtained about 3000 individual earthquakes and  
 221 about 500 SSEs in our calculations when the faults hosted both earthquakes and SSEs.

222 We identify five different domains of fault slip behavior (Fig. 2). For small faults  
 223 ( $L \ll L_{nuc}$ ), there is a damped domain in which the fault experiences no events at all  
 224 as the fault length is too small for any type of instability to grow. For long faults ( $L \gg$   
 225  $L_{nuc}$ ) with strongly rate-weakening properties ( $a/b < 0.5$ ), we observe periodic earth-  
 226 quakes, similar as in a case with no geometrical complexity. This is perfectly normal as  
 227 both our faults are flat and the longer they are, the larger the portion that is left unaf-  
 228 fected by the geometrical complexity (i.e. if the faults are long, their edges are indepen-  
 229 dent and dominate the general behavior of slip, reducing this setting to a case with no  
 230 geometrical complexity). For mildly rate-weakening faults ( $1 > a/b > 0.6$ ) and what-  
 231 ever the length of the fault, we observe a complex behavior with a mixture of slow and  
 232 rapid slip for fault sizes between 1 and 4 times the nucleation length and only complex  
 233 earthquakes (partial ruptures, aperiodic events, variable after slip) for longer faults. That

234 is, although the length over which we observe slow slip events is increased compared to  
 235 the case where there is no additional fault, we are still limited by the nucleation length  
 236 scale. Therefore like in other studies, we will require another mechanism. This can just be  
 237 low effective normal stress, additional frictional mechanisms like dilatant strengthening or  
 238 even stronger geometrical complexities. The domain where both slow and fast earthquake  
 239 coexist, shrinks when the distance between the faults is increased. All this put together  
 240 confirms our intuition that stress perturbations from one fault to another help modulate the  
 241 mode of slip along faults.

## 242 **4.2 Scaling**

243 Geodetic and seismological observations in nature suggest two different scaling rela-  
 244 tionships for moment of slow slip on one side and rapid, dynamic slip events on the other  
 245 side [*Ide et al., 2007; Peng and Gomberg, 2010*]. Considering the statistics of slip events  
 246 produced by our model, we also find that the moment of both seismic and aseismic events  
 247 modeled by rate and state friction law follows two different scaling laws as observed in  
 248 nature (Fig. 3). Because we conducted our calculations in 2D, the moment of a dynamic  
 249 slip event should scale with its duration squared:  $M \propto T^2$ . This scaling emerges naturally  
 250 from our conceptual model without imposing any complexity in the spatial variation of  
 251 frictional properties. If we do not include any geometrical complexity, periodic, identical  
 252 earthquakes are observed impeding our ability to observe any potential scaling. Although  
 253 we do not preclude the possibility that other models, that have produced SSE's and earth-  
 254 quakes, also reproduce such scaling laws, geometrical complexities give rise to a wide  
 255 range of modes of slip and the resulting events obey similar scaling laws as in nature.

256 We note the moment of our simulated events clearly depends on the ratio of consti-  
 257 tutive parameters  $a/b$ . Since the nucleation length  $L_{nuc}$  increases with  $a/b$  and since we  
 258 compare models with non-dimensionalised fault length, the real length of the fault,  $L$ , also  
 259 increases when  $a/b \rightarrow 1$ , leading to bigger moment release and longer duration for events.  
 260 To verify the robustness of this scaling law, we changed the maximum slip velocity cri-  
 261 teria used to distinguish SSEs and earthquakes by one order of magnitude. This does not  
 262 change the observed scaling.

263 Another interesting feature that emerges from our calculations is that the moment  
 264 of the nucleation phase of earthquakes also follows the same linear scaling with duration

265 as slow-slip events. However, we cannot argue that this similarity in scaling may be pre-  
 266 served in 3D. We finally notice that by adding the nucleation and after-slip moment of  
 267 earthquakes, the clear scaling distinction between earthquakes and SSEs starts vanishing  
 268 (see Figure S1 in Supp. Mat.). This observation is in favor of a continuum of modes of  
 269 slip ranging from slow to rapid, dynamic slip.

270 We can find some physical intuition about this relative scaling between SSEs and  
 271 earthquakes in the temporal evolution of rupture length and slip for each event (Fig. 4).  
 272 For earthquakes, the average growth of both rupture length and slip are linear with event  
 273 duration, independent of  $a/b$ , hence independent of the actual length of the fault as we  
 274 non-dimensionalised length scales by  $L_{nuc}$ . As a consequence, seismic moment grows  
 275 quadratically with event duration. In other words, earthquakes propagate as an expanding  
 276 crack: slip and rupture length are proportional to each other.

277 For SSEs, however, the temporal evolution of slip and rupture length shows a clear  
 278 dependence on the fault length. For a given  $a/b$ , final rupture length is constant i.e. it is  
 279 independent of event duration. However, slip grows linearly with duration. If we now in-  
 280 crease the fault length (i.e. increase  $a/b$ ), the accumulated slip decreases (compared to  
 281 the low  $a/b$  case) while the final rupture length increases. These two effects exactly coun-  
 282 terbalance each other, such that the final moment scales linearly with duration and is in-  
 283 dependent of fault length (i.e. for different  $a/b$ ). This highlights an interesting fact that  
 284 SSEs are not necessarily self-similar in our calculations.

285 Finally, we observe that the moment of the nucleation phase scales linearly with its  
 286 duration. The evolution of slip and rupture length for the nucleation phase is scale inde-  
 287 pendent contrary to SSEs. Slip and final rupture length for nucleation phases evolve, in-  
 288 dividually, with the square root of the event duration, which might point to a significant  
 289 difference between these processes.

### 290 **4.3 Stress drop**

291 Interestingly, static stress drops of both slow and rapid slip events in our model are  
 292 comparable (see Figure S4 in Supp. Mat.). We evaluate this parameter in three different  
 293 ways following *Noda et al.* [2013] (see Supp. Mat. for more details). Regardless of the  
 294 method, stress drops of SSEs and earthquakes are of similar order of magnitude. Earth-  
 295 quake stress drops are on an average about twice as large as those for SSEs. This is not

296 completely in agreement with observations where SSEs stress drop is generally 1 or 2 or-  
297 ders of magnitude smaller than for earthquakes [Gao *et al.*, 2012]. However, it has also  
298 been shown that earthquake stress drops can vary by several orders of magnitude [Goebel  
299 *et al.*, 2015]. Finally, and as expected, the stress drop scales with the moment of individ-  
300 ual earthquakes and SSEs. Such observation emphasises the relative importance of slow  
301 events in the stress/energy budget of active faults.

## 302 **5 Conclusion**

303 We have shown that one simple geometrical complexity (two overlapping faults) can  
304 naturally result in a complex seismic cycle (with SSEs, earthquakes, partial ruptures etc.),  
305 without appealing to complex friction rheology on the fault. We believe that geometry of  
306 fault systems, that have been shown to control the dynamics of ordinary earthquakes [Lay  
307 *and Kanamori*, 1981], are also a primary cause of the source of complexity in the seismic  
308 cycle.

309 In recent years, many models have attempted to explain the nearly ubiquitous pres-  
310 ence of slow-slip events in subduction zone. Current models using rate and state friction  
311 can only produce slow and fast dynamics in a very narrow range of parameters. Exten-  
312 sion of this range required considering additional competing frictional mechanisms. Our  
313 work here suggests that complex stress interaction due to geometric complexity of faults  
314 could also act as a complementary mechanism to enhance the presence of slow slip in  
315 models. This work is an exploratory work on the role of fault geometric complexities in  
316 an earthquake cycle. We think that the role of fault geometry in earthquake cycle mod-  
317 els has been under-emphasised compared to the role of friction laws in earthquake cy-  
318 cle modelling probably because of the inherent computational limitation of modelling on  
319 non-planar geometries. We argue that a unified model that would explain all observations  
320 needs to account for geometric segmentation and/or the non-planar nature of active faults  
321 as this is a first-order and well documented feature that results in a spatiotemporally inho-  
322 mogeneous stress accumulation rate [Mitsui *and Hirahara*, 2006; Matsuzawa *et al.*, 2013;  
323 Li *and Liu*, 2016]. As this work shows, the simplest of geometrical complexity can lead to  
324 very complex modes of slip on a fault network.

325 **Acknowledgments**

326 Numerical computations were performed on the S-CAPAD platform, IGP, France. P. R.  
327 and H. S. B. are grateful to Leslie Greengard and Zydrunas Gimbutas for the FMMLIB2D  
328 library. This article benefited from discussions with Robert Viesca and Pierre Dublanchet.  
329 P. R. acknowledges the GPX program, funded by the French National Research Agency  
330 (ANR), CGG, TOTAL and Schlumberger, for his PhD fellowship. This work received  
331 funding from the European Research Council (ERC) under the European Union's Hori-  
332 zon 2020 research and innovation program (Geo-4D project, grant agreement 758210).  
333 The software developed for the paper and all the relevant data is available permanently at  
334 <http://www.geologie.ens.fr/~bhat/romanetGRL2018/>.

**References**

- 360  
361 Araki, E., D. M. Saffer, A. J. Kopf, L. M. Wallace, T. Kimura, Y. Machida, S. Ide,  
362 E. Davis, I. Expedition, et al. (2017), Recurring and triggered slow-slip events near the  
363 trench at the Nankai Trough subduction megathrust, *Science*, *356*(6343), 1157–1160, doi:  
364 10.1126/science.aan3120.
- 365 Bilham, R., H. Ozener, D. Mencin, A. Dogru, S. Ergintav, Z. Cakir, A. Aytun, B. Aktug,  
366 O. Yilmaz, W. Johnson, and G. Mattioli (2016), Surface creep on the North Anatolian  
367 fault at Isetpasa, Turkey, 1944–2016, *J. Geophys. Res.*, *121*(10), 7409–7431, doi:10.  
368 1002/2016JB013394.
- 369 Brodsky, E. E., and T. Lay (2014), Recognizing foreshocks from the 1 April 2014 Chile  
370 Earthquake, *Science*, *344*(6185), 700–702, doi:10.1126/science.1255202.
- 371 Candela, T., F. Renard, Y. Klinger, K. Mair, J. Schmittbuhl, and E. E. Brodsky (2012),  
372 Roughness of fault surfaces over nine decades of length scales, *J. Geophys. Res.*, *117*,  
373 B08,409, doi:10.1029/2011JB009041.
- 374 Carrier, J., L. Greengard, and V. Rokhlin (1988), A fast adaptive multipole algorithm for  
375 particle simulations, *SIAM J. Sci. Stat. Comput.*, *9*(4), 669–686.
- 376 Das, S., and A. B. Watts (2009), Effect of subducting seafloor topography on the rupture  
377 characteristics of great subduction zone earthquakes, in *Subduction Zone Geodynam-*  
378 *ics*, edited by S. E. Lallemand and F. Funiciello, pp. 103–118, Springer-Verlag, Berlin-  
379 Heidelberg, doi:10.1007/978-3-540-87974-9.
- 380 Dieterich, J. H. (1978), Time-dependent friction and the mechanics of stick-slip, *Pure*  
381 *Appl. Geophys.*, *116*(4-5), 790–806.
- 382 Dieterich, J. H. (1992), Earthquake nucleation on faults with rate- and state-dependent  
383 strength, *Tectonophysics*, *211*(1-4), 115–134.
- 384 Douglas, A., J. Beavan, L. Wallace, and J. Townend (2005), Slow slip on the northern  
385 Hikurangi subduction interface, New Zealand, *Geophys. Res. Lett.*, *32*(16), doi:10.1029/  
386 2005GL023607.
- 387 Dragert, H., K. Wang, and T. S. James (2001), A silent slip event on the deeper Cascadia  
388 subduction interface, *Science*, *292*(5521), 1525–1528.
- 389 Dragert, H., K. Wang, and G. Rogers (2004), Geodetic and seismic signatures of episodic  
390 tremor and slip in the northern Cascadia subduction zone, *Earth Planets Space*, *56*(12),  
391 1143–1150.

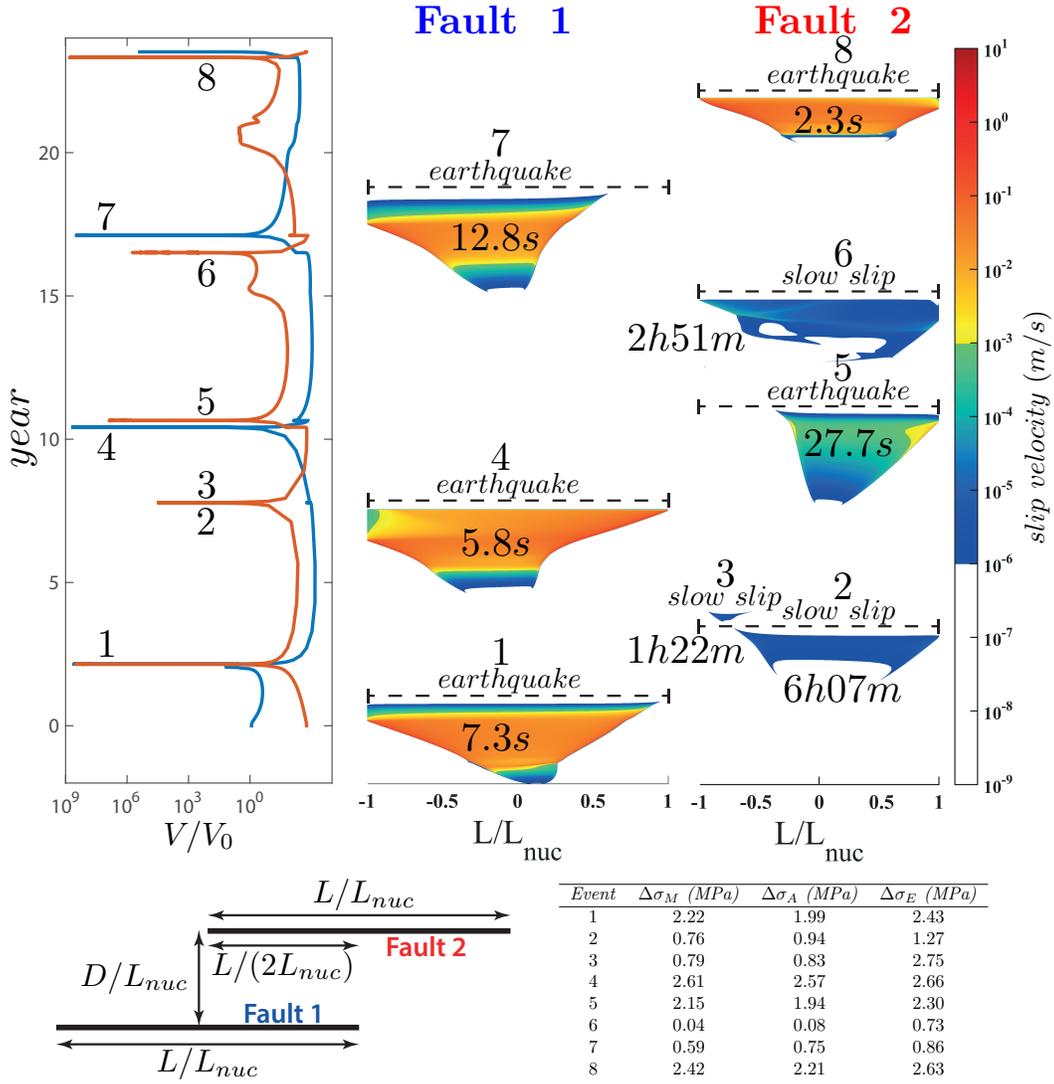
- 392 Franco, S., V. Kostoglodov, K. Larson, V. Manea, M. Manea, and J. Santiago (2005),  
393 Propagation of the 2001-2002 silent earthquake and interplate coupling in the Oaxaca  
394 subduction zone, Mexico, *Earth Planets Space*, 57(10), 973–985.
- 395 Gao, H., D. A. Schmidt, and R. J. Weldon (2012), Scaling relationships of source pa-  
396 rameters for slow slip events, *Bull. Seism. Soc. Am.*, 102(1), 352–360, doi:10.1785/  
397 0120110096.
- 398 Goebel, T. H. W., E. Hauksson, P. M. Shearer, and J.-P. Ampuero (2015), Stress-drop  
399 heterogeneity within tectonically complex regions: a case study of San Geronimo Pass,  
400 southern California, *Geophys. J. Int.*, 202(1), 514–528, doi:10.1093/gji/ggv160.
- 401 Gomberg, J., A. Wech, K. Creager, K. Obara, and D. Agnew (2016), Reconsidering earth-  
402 quake scaling, *Geophys. Res. Lett.*, 43(12), 6243–6251, doi:10.1002/2016GL069967.
- 403 Greengard, L., and V. Rokhlin (1987), A fast algorithm for particle simulations, *J. Com-  
404 put. Phys.*, 73(2), 325–348.
- 405 Gualandi, A., C. Nichele, E. Serpelloni, L. Chiaraluce, L. Anderlini, D. Latorre, M. Be-  
406 lardinelli, and J.-P. Avouac (2017), Aseismic deformation associated with an earth-  
407 quake swarm in the northern Apennines (Italy), *Geophys. Res. Lett.*, doi:10.1002/  
408 2017GL073687.
- 409 Hicks, S. P., and A. Rietbrock (2015), Seismic slip on an upper-plate normal fault during  
410 a large subduction megathrust rupture, *Nature Geosci.*, 8(12), 955–960, doi:10.1038/  
411 NGE02585.
- 412 Hirose, H., K. Hirahara, F. Kimata, N. Fujii, and S. Miyazaki (1999), A slow thrust slip  
413 event following the two 1996 Hyuganada earthquakes beneath the Bungo Channel, south-  
414 west Japan, *Geophys. Res. Lett.*, 26(21), 3237–3240.
- 415 Hubbard, J., S. Barbot, E. M. Hill, and P. Tapponnier (2015), Coseismic slip on shallow  
416 décollement megathrusts: implications for seismic and tsunami hazard, *Earth-Science  
417 Reviews*, 141, 45–55, doi:10.1016/j.earscirev.2014.11.003.
- 418 Ide, S. (2014), Modeling fast and slow earthquakes at various scales, *Proceedings of the  
419 Japan Academy. Series B, Physical and biological sciences*, 90(8), 259, doi:10.2183/pjab.  
420 90.259.
- 421 Ide, S., G. C. Beroza, D. R. Shelly, and T. Uchide (2007), A scaling law for slow earth-  
422 quakes, *Nature*, 447(7140), 76–79, doi:10.1038/nature05780.
- 423 Ide, S., K. Imanishi, Y. Yoshida, G. C. Beroza, and D. R. Shelly (2008), Bridging the gap  
424 between seismically and geodetically detected slow earthquakes, *Geophys. Res. Lett.*,

- 425 35(10), L10,305, doi:10.1029/2008GL034014.
- 426 Igarashi, T., T. Matsuzawa, and A. Hasegawa (2003), Repeating earthquakes and inter-  
427 plate aseismic slip in the northeastern japan subduction zone, *J. Geophys. Res.*, *108*(B5),  
428 2249, doi:10.1029/2002JB001920.
- 429 Itaba, S., and R. Ando (2011), A slow slip event triggered by teleseismic surface waves,  
430 *Geophys. Res. Lett.*, *38*(21), L21,306, doi:10.1029/2011GL049593,2011.
- 431 Ito, Y., R. Hino, M. Kido, H. Fujimoto, Y. Osada, D. Inazu, Y. Ohta, T. Iinuma,  
432 M. Ohzono, S. Miura, M. Mishina, K. Suzuki, T. Tsuji, and J. Ashi (2013), Episodic  
433 slow slip events in the japan subduction zone before the 2011 tohoku-oki earthquake,  
434 *Tectonophysics*, *600*, 14–26, doi:10.1016/j.tecto.2012.08.022.
- 435 Jolivet, R., C. Lasserre, M.-P. Doin, G. Peltzer, J.-P. Avouac, J. Sun, and R. Dailu (2013),  
436 Spatio-temporal evolution of aseismic slip along the haiyuan fault, china: Implications  
437 for fault frictional properties, *Earth Planet. Sc. Lett.*, *377*, 23–33, doi:10.1016/j.epsl.  
438 2013.07.020.
- 439 Kato, A., K. Obara, T. Igarashi, H. Tsuruoka, S. Nakagawa, and N. Hirata (2012), Propa-  
440 gation of Slow Slip Leading Up to the 2011 Mw 9.0 Tohoku-Oki Earthquake, *Science*,  
441 *335*(6069), 705–708, doi:10.1126/science.1215141.
- 442 Kato, A., T. Igarashi, and K. Obara (2014), Detection of a hidden boso slow slip event  
443 immediately after the 2011 mw 9.0 tohoku-oki earthquake, japan, *Geophysical Research*  
444 *Letters*, *41*(16), 5868–5874, doi:10.1002/2014GL061053.
- 445 Lay, T., and H. Kanamori (1981), An asperity model of large earthquake sequences, in  
446 *Earthquake Prediction, an International Review, Maurice Ewing Series*, vol. IV, edited by  
447 D. W. Simpson and P. Richards, pp. 579–592, AGU, Washington, D. C., doi:10.1029/  
448 ME004p0579.
- 449 Li, D., and Y. Liu (2016), Spatiotemporal evolution of slow slip events in a nonplanar  
450 fault model for northern cascadia subduction zone, *J. Geophys. Res.*, *121*(9), 6828–6845,  
451 doi:10.1002/2016JB012857.
- 452 Liu, Y., and J. R. Rice (2005), Aseismic slip transients emerge spontaneously in 3d rate  
453 and state modeling of subduction earthquake sequences, *J. Geophys. Res.*, *110*, B08,307,  
454 doi:10.1029/2004JB003424.
- 455 Liu, Y., and J. R. Rice (2007), Spontaneous and triggered aseismic deformation tran-  
456 sients in a subduction fault model, *J. Geophys. Res.*, *112*, B09,404, doi:10.1029/  
457 2007JB004930.

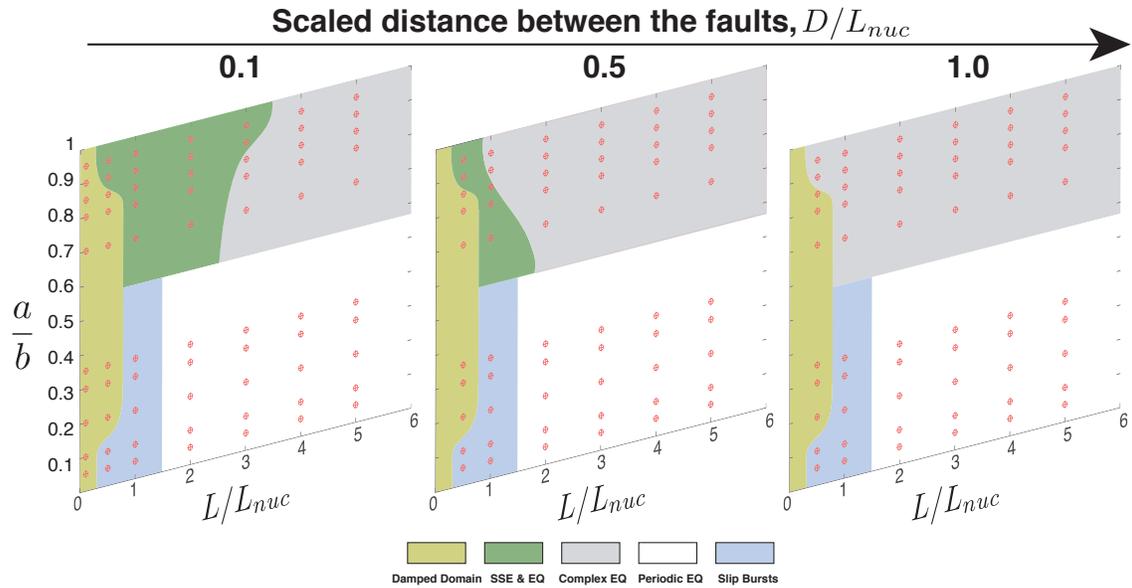
- 458 Lowry, A. R., K. M. Larson, V. Kostoglodov, and R. Bilham (2001), Transient fault slip  
459 in guerrero, southern mexico, *Geophys. Res. Lett.*, *28*(19), 3753–3756, doi:10.1029/  
460 2001GL013238.
- 461 Matsuzawa, T., B. Shibazaki, K. Obara, and H. Hirose (2013), Comprehensive model of  
462 short-and long-term slow slip events in the shikoku region of japan, incorporating a  
463 realistic plate configuration, *Geophys. Res. Lett.*, *40*(19), 5125–5130, doi:10.1002/grl.  
464 51006.
- 465 Mavrommatis, A. P., P. Segall, N. Uchida, and K. M. Johnson (2015), Long-term accel-  
466 eration of aseismic slip preceding the Mw 9 Tohoku-oki earthquake: Constraints from  
467 repeating earthquakes, *Geophys. Res. Lett.*, *42*, 9717–9725, doi:10.1002/2015GL066069.
- 468 Mitsui, N., and K. Hirahara (2006), Slow slip events controlled by the slab dip and its  
469 lateral change along a trench, *Earth Planet. Sc. Lett.*, *245*(1), 344–358, doi:10.1016/j.  
470 epsl.2006.03.001.
- 471 Murray, J. R., and P. Segall (2005), Spatiotemporal evolution of a transient slip event on  
472 the san andreas fault near parkfield, california, *J. Geophys. Res.*, *110*(B9), doi:10.1029/  
473 2005JB003651.
- 474 Noda, H., N. Lapusta, and H. Kanamori (2013), Comparison of average stress drop mea-  
475 sures for ruptures with heterogeneous stress change and implications for earthquake  
476 physics, *Geophys. J. Int.*, *193*, 1691—1712, doi:10.1093/gji/ggt074.
- 477 Obara, K., and A. Kato (2016), Connecting slow earthquakes to huge earthquakes, *Sci-*  
478 *ence*, *353*(6296), 253–257, doi:10.1126/science.aaf1512.
- 479 Peng, Z., and J. Gomberg (2010), An integrated perspective of the continuum between  
480 earthquakes and slow-slip phenomena, *Nature Geoscience*, *3*(9), 599–607, doi:10.1038/  
481 ngeo940.
- 482 Radiguet, M., H. Perfettini, N. Cotte, A. Gualandi, B. Valette, V. Kostoglodov,  
483 T. Lhomme, A. Walpersdorf, E. Cabral Cano, and M. Campillo (2016), Triggering of  
484 the 2014 mw7.3 papanoa earthquake by a slow slip event in guerrero, mexico, *Nature*  
485 *Geoscience*, *9*, 829–833, doi:10.1038/NGEO2817.
- 486 Rice, J. R. (1983), Constitutive relations for fault slip and earthquake instabilities, *Pure*  
487 *Appl. Geophys.*, *121*(3), 443–475.
- 488 Rice, J. R. (1993), Spatio-temporal complexity of slip on a fault, *J. Geophys. Res.*, *98*(B6),  
489 9885–9907.

- 490 Rogers, G., and H. Dragert (2003), Episodic tremor and slip on the Cascadia subduction  
491 zone: The chatter of silent slip, *Science*, *300*(5627), 1942–1943.
- 492 Rousset, B., R. Jolivet, M. Simons, C. Lasserre, B. Riel, P. Milillo, Z. Çakir, and F. Re-  
493 nard (2016), An aseismic slip transient on the north anatolian fault, *Geophys. Res. Lett.*,  
494 *43*(7), 3254–3262, doi:10.1002/2016GL068250.
- 495 Rubin, A., and J.-P. Ampuero (2005), Earthquake nucleation on (aging) rate and state  
496 faults, *J. Geophys. Res.*, *110*, B11,312, doi:10.1029/2005JB003686.
- 497 Rubin, A. M. (2008), Episodic slow slip events and rate-and-state friction, *J. Geophys.*  
498 *Res.*, *113*, B11,414, doi:10.1029/2008JB005642.
- 499 Ruina, A. (1983), Slip instability and state variable friction laws, *J. Geophys. Res.*, *88*(10),  
500 359–370.
- 501 Ruiz, S., M. Metois, A. Fuenzalida, J. Ruiz, F. Leyton, R. Grandin, C. Vigny,  
502 R. Madariaga, and J. Campos (2014), Intense foreshocks and a slow slip event pre-  
503 ceded the 2014 iquique mw 8.1 earthquake, *Science*, *345*(6201), 1165–1169, doi:  
504 10.1126/science.1256074.
- 505 Saffer, D. M., and L. M. Wallace (2015), The frictional, hydrologic, metamorphic  
506 and thermal habitat of shallow slow earthquakes, *Nature Geoscience*, doi:10.1038/  
507 NGE02490.
- 508 Schwartz, S. Y., and J. M. Rokoşky (2007), Slow slip events and seismic tremor  
509 at circum-pacific subduction zones, *Rev. Geophys.*, *45*(3), 1–32, doi:10.1029/  
510 2006RG000208.
- 511 Segall, P., and A. M. Bradley (2012), Slow-slip evolves into megathrust earthquakes  
512 in 2d numerical simulations, *Geophys. Res. Lett.*, *39*(18), L18,308, doi:10.1029/  
513 2012GL052811.
- 514 Segall, P., and J. R. Rice (1995), Dilatancy, compaction, and slip instability of a fluid-  
515 infiltrated fault, *J. Geophys. Res.*, *100*(B11), 22,155–22,171.
- 516 Segall, P., A. M. Rubin, A. M. Bradley, and J. R. Rice (2010), Dilatant strengthening  
517 as a mechanism for slow slip events, *J. Geophys. Res.*, *115*, B12,305, doi:10.1029/  
518 2010JB007449,.
- 519 Sekine, S., H. Hirose, and K. Obara (2010), Along-strike variations in short-term slow slip  
520 events in the southwest japan subduction zone, *J. Geophys. Res.*, *115*(B9), doi:10.1029/  
521 2008JB006059.

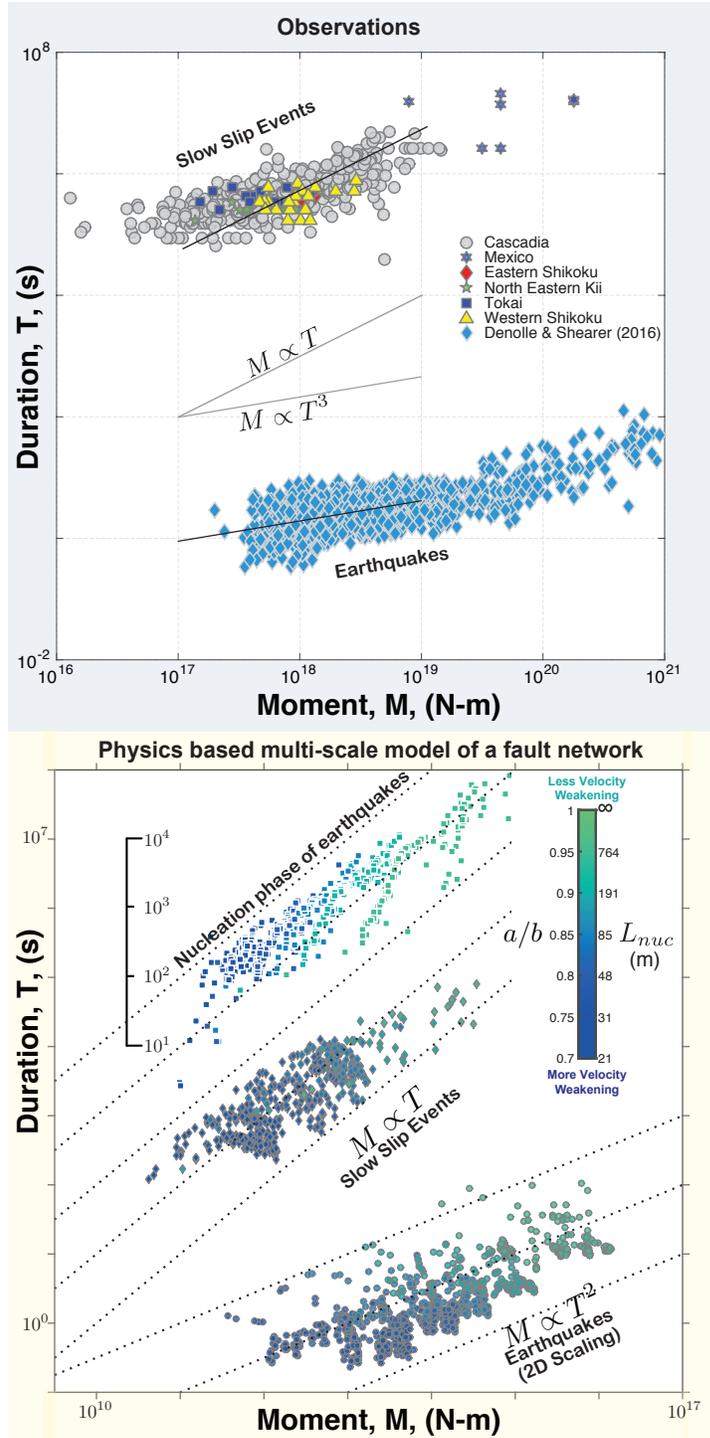
- 522 Thomas, A. M., R. M. Nadeau, and R. Bürgmann (2009), Tremor-tide correlations and  
523 near-lithostatic pore pressure on the deep San Andreas fault, *Nature*, *462*(7276), 1048–  
524 1051, doi:10.1038/nature08654.
- 525 Vallee, M., J.-M. Nocquet, J. Battaglia, Y. Font, M. Segovia, M. Regnier, P. Mothes,  
526 P. Jarrin, D. Cisneros, S. Vaca, et al. (2013), Intense interface seismicity triggered by a  
527 shallow slow slip event in the central ecuador subduction zone, *J. Geophys. Res.*, *118*(6),  
528 2965–2981, doi:10.1002/jgrb.50216.
- 529 Viesca, R. C. (2016), Stable and unstable development of an interfacial sliding instability,  
530 *Phys. Rev. E*, *93*(6), 060,202, doi:10.1103/PhysRevE.93.060202.
- 531 Wallace, L. M., S. C. Webb, Y. Ito, K. Mochizuki, R. Hino, S. Henrys, S. Y. Schwartz,  
532 and A. F. Sheehan (2016), Slow slip near the trench at the hikurangi subduction zone,  
533 new zealand, *Science*, *352*(6286), 701–704.
- 534 Wallace, L. M., Y. Kaneko, S. Hreinsdóttir, I. Hamling, Z. Peng, N. Bartlow,  
535 E. D’Anastasio, and B. Fry (2017), Large-scale dynamic triggering of shallow slow slip  
536 enhanced by overlying sedimentary wedge, *Nature Geoscience*, doi:10.1038/NGEO3021.
- 537 Zigone, D., D. Rivet, M. Radiguet, M. Campillo, C. Voisin, N. Cotte, A. Walpersdorf,  
538 N. M. Shapiro, G. Cougoulat, P. Roux, et al. (2012), Triggering of tremors and slow  
539 slip event in guerrero, mexico, by the 2010 mw 8.8 maule, chile, earthquake, *J. Geo-*  
540 *phys. Res.*, *117*, B09,304, doi:10.1029/2012JB009160.



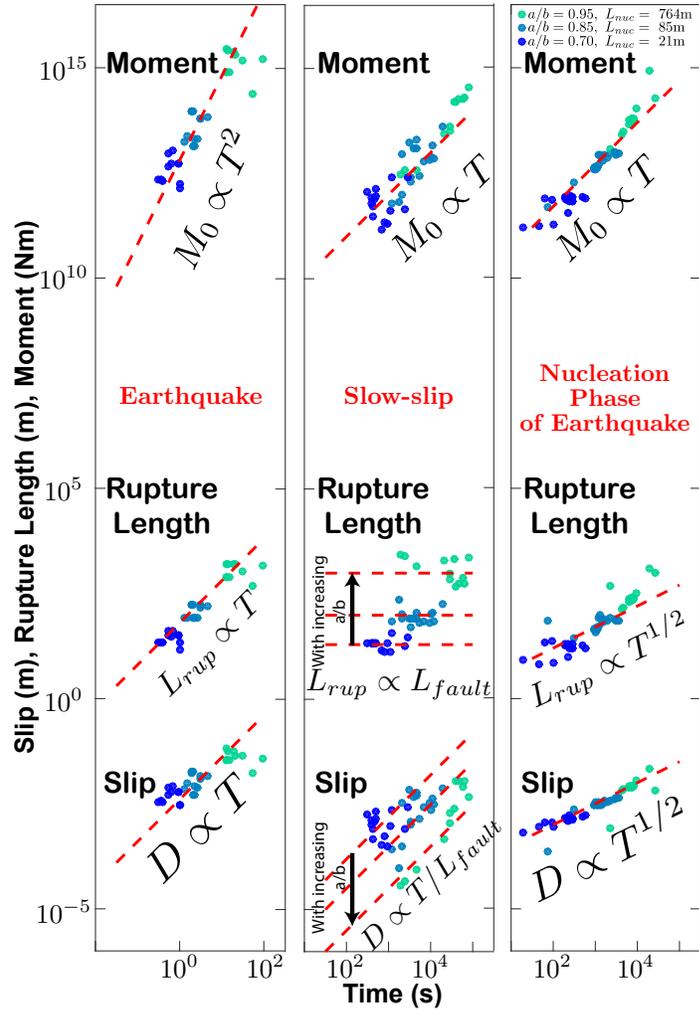
335 **Figure 1.** Example of a calculation that gives rise to complex slip behaviour on faults. Here  $L/L_{nuc} = 2$ ,  
 336  $D/L_{nuc} = 0.1$  and  $a/b = 0.9$ . To avoid any artefact from initial conditions, the first 10 events of the simu-  
 337 lation were removed. Left panel shows the maximum slip velocity for fault 1 (blue) and fault 2 (red). Right  
 338 panel represents the space-time evolution of slip velocity on the faults. The highlighted duration of events  
 339 corresponds respectively for earthquakes and slow events to the time when the slip velocity exceeds 1mm/s  
 340 or  $1\mu\text{m/s}$  for the first time to the time when it decelerates below 1mm/s or  $1\mu\text{m/s}$ . Bottom panel gives the  
 341 geometry used for this example. Events 2,3 and 6 are slow-slip events. Events 1, 4, 5, 7 and 8 are earthquakes.  
 342 Event 5 and 7 are small earthquakes that did not rupture the entire fault. Event 1 and 7 clearly show after-  
 343 slip contrary to events 4 and 8. The table lists the seismological ( $\Delta\sigma_M$ ), spatially averaged ( $\Delta\sigma_A$ ) and slip  
 344 averaged ( $\Delta\sigma_E$ ) stress drops for the events.



345 **Figure 2.** Phase diagram showing the evolution of mode of slip along the 2 fault system given the distance  
 346 between the faults. This figure includes a broader set of simulations in comparison to the paper. *Damped*  
 347 *domain* is a domain within which the fault experiences no event at all. *SSE & EQ* is the domain of coexistence  
 348 of both slow events and earthquakes. *Complex EQ* is a domain within which we get only earthquakes but with  
 349 spatio-temporal complexities. *Periodic EQ* is a domain within which earthquakes are periodically rupturing  
 350 the entire fault. And finally, *Slip Bursts* is a domain within which the entire fault is destabilized at the same  
 351 time, there is no propagation of the rupture. This corresponds for small faults compared to the nucleation  
 352 lengthscale and small  $a/b$ . This domain is called the no-healing regime [Rubin and Ampuero, 2005].

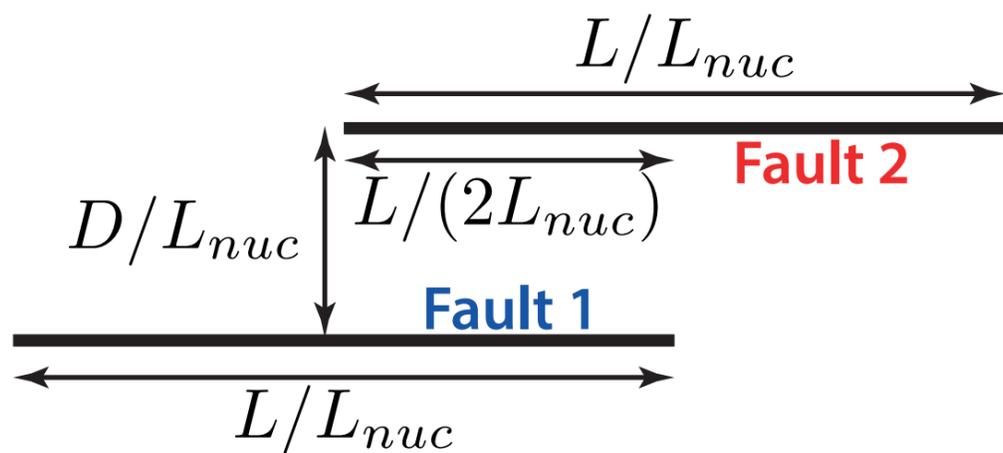
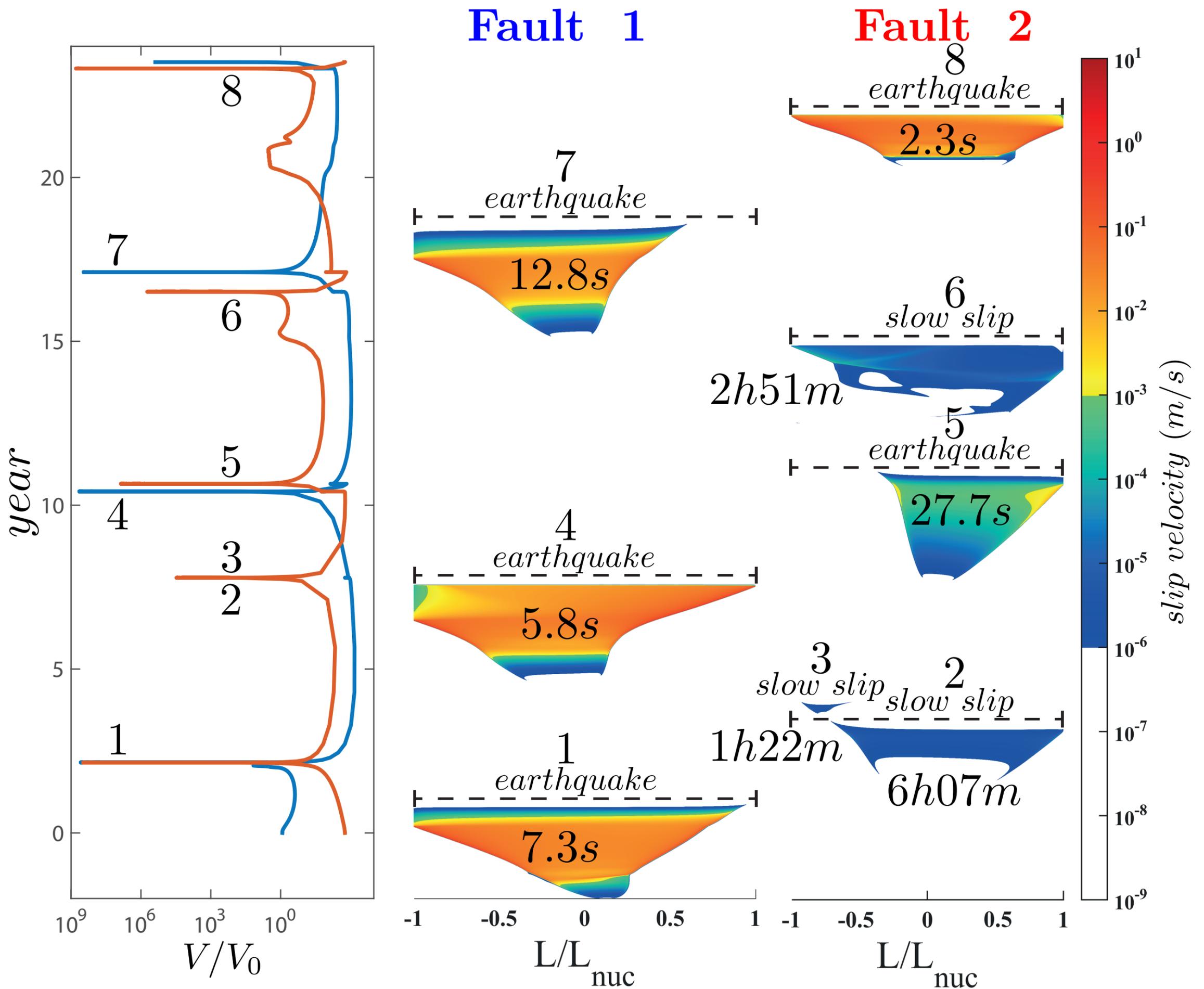


353 **Figure 3.** Comparison of the scaling law for observational data [Sekine *et al.*, 2010; Gao *et al.*, 2012;  
 354 Gombert *et al.*, 2016] (top panel) and from our all our calculations (bottom panel). We only used the seis-  
 355 mic moment of the dynamic part of an earthquake. The original scaling [Ide *et al.*, 2007] also included data  
 356 from tremors, very low frequency earthquakes and low frequency earthquake. However because we are not  
 357 reproducing any of these events, we cut the data to show only slow slip events.



358 **Figure 4.** Final moment, slip and rupture length with time for slow-slip events, earthquakes and  
 359 nucleation phase of earthquakes.

**Figure1.**



| Event | $\Delta\sigma_M$ (MPa) | $\Delta\sigma_A$ (MPa) | $\Delta\sigma_E$ (MPa) |
|-------|------------------------|------------------------|------------------------|
| 1     | 2.22                   | 1.99                   | 2.43                   |
| 2     | 0.76                   | 0.94                   | 1.27                   |
| 3     | 0.79                   | 0.83                   | 2.75                   |
| 4     | 2.61                   | 2.57                   | 2.66                   |
| 5     | 2.15                   | 1.94                   | 2.30                   |
| 6     | 0.04                   | 0.08                   | 0.73                   |
| 7     | 0.59                   | 0.75                   | 0.86                   |
| 8     | 2.42                   | 2.21                   | 2.63                   |

Figure2.

# Scaled distance between the faults, $D/L_{nuc}$

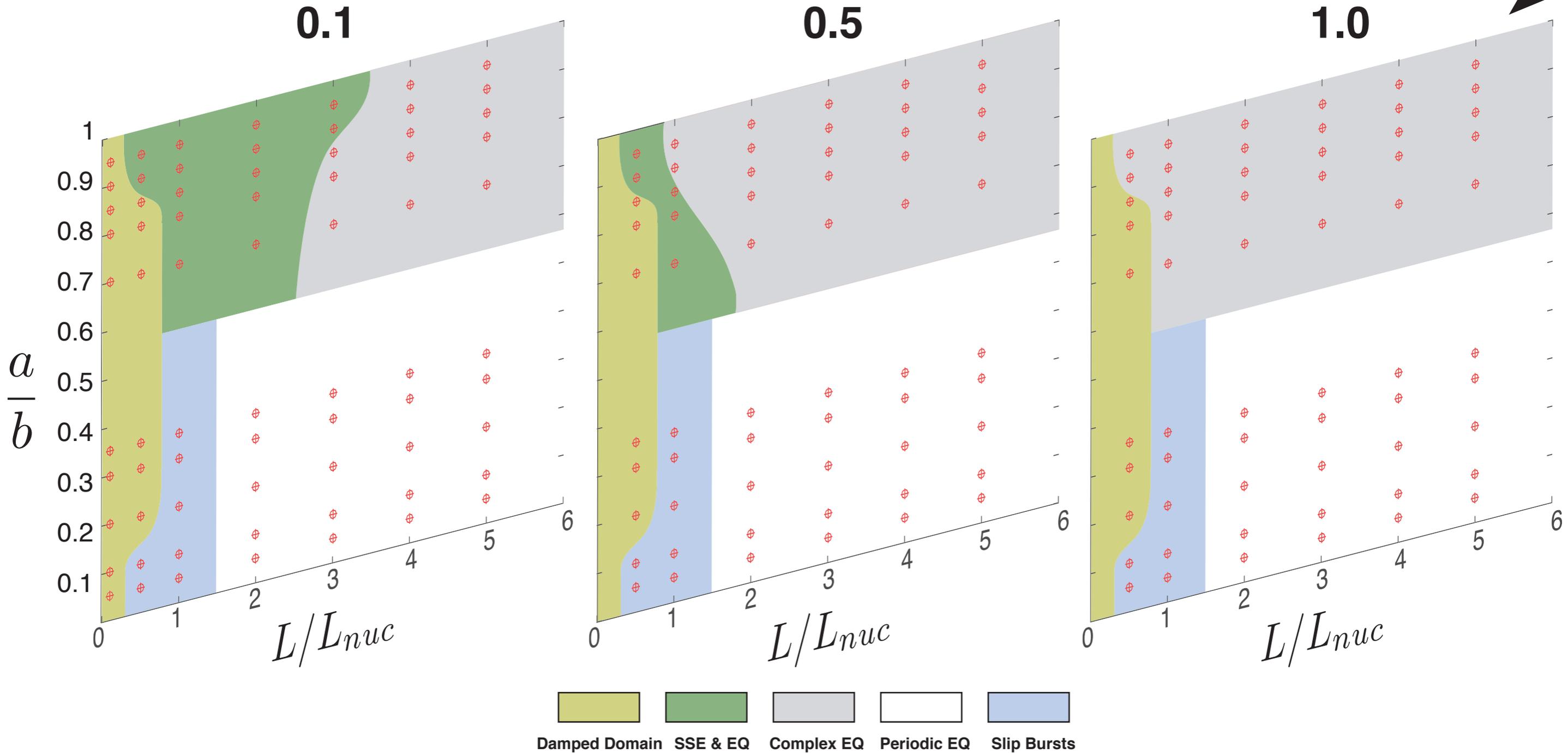
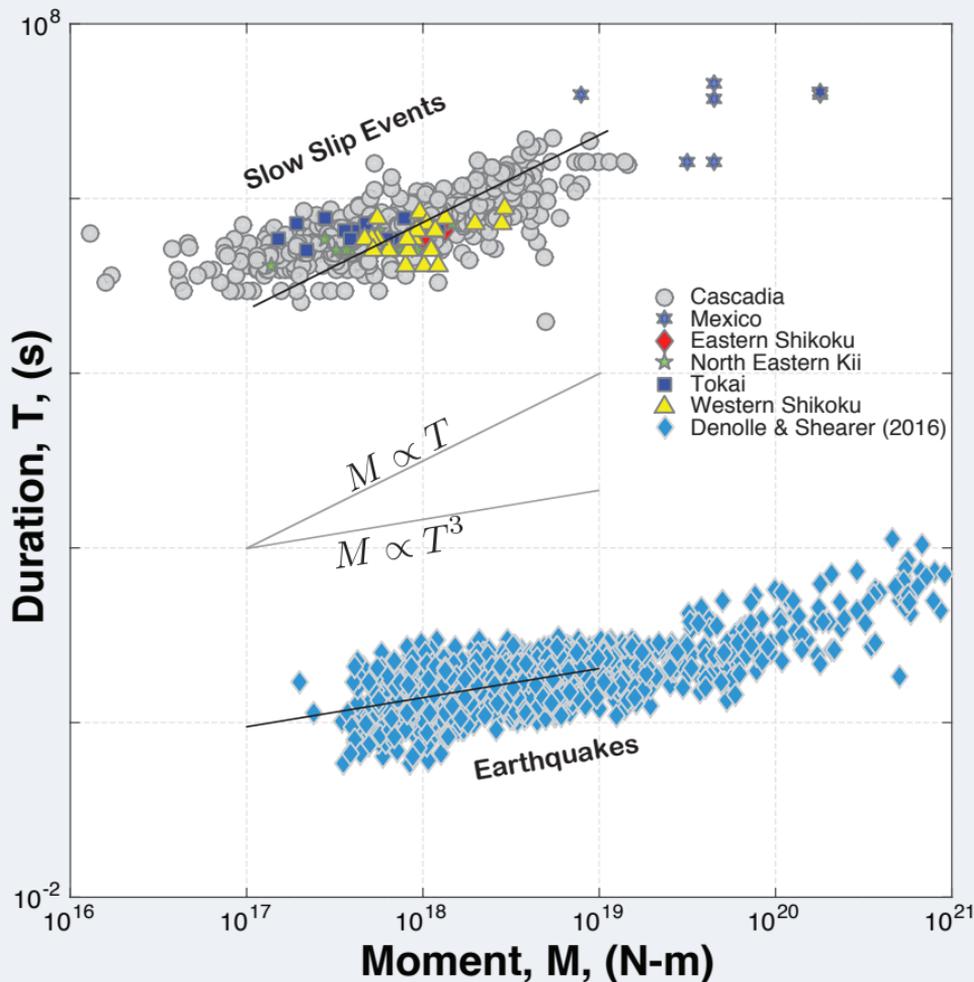


Figure3.

### Observations



### Physics based multi-scale model of a fault network

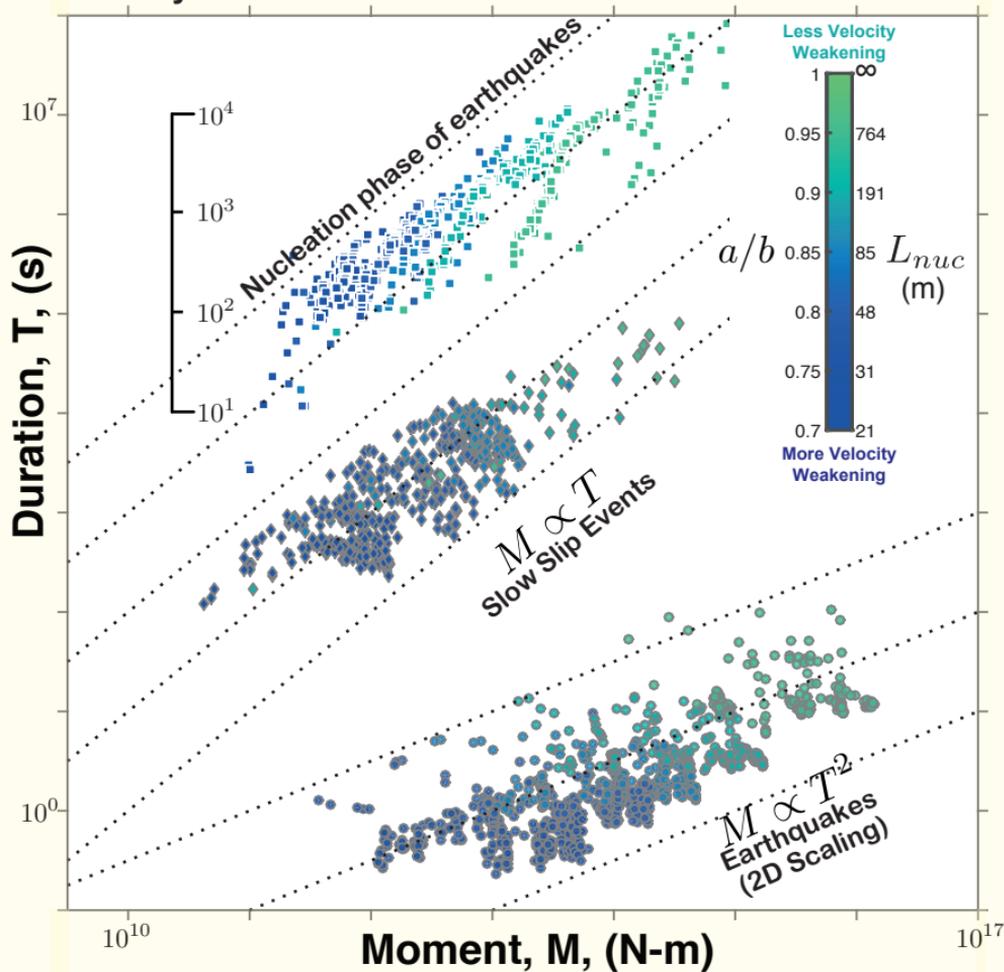


Figure4.

