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Episodicity and Migration of Low Frequency Earthquakes modeled with Fast Fluid Pressure Transients in the Permeable Subduction Interface

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1 **Episodicity and Migration of Low Frequency**
2 **Earthquakes modeled with Fast Fluid Pressure**
3 **Transients in the Permeable Subduction Interface**

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8 **Key Points:**

- 9 • We model fluid pressure diffusion in a channel along the subduction interface in
10 which the permeability varies through a valving mechanism
- 11 • Tremor is generated as valves break open and interactions between valves through
12 fluid pressure transients shape realistic event patterns
- 13 • The input fluid flux in the fault controls seismic activity, opening possibilities to
14 diagnose hydraulic conditions using LFE patterns

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Abstract

In many subduction zones, the plate interface hosts intermittent, low-frequency, low-magnitude seismic tremor and low-frequency earthquakes (LFEs). Seismic activity clusters in episodic bursts that migrate along the fault zone in complex ways. Geological structures in fossil tremor source regions testify to large and pervasive variations of fluid pressure and permeability. Here, we explore the potential of fluid pressure transients in a permeable subduction interface to trigger seismic sources and induce interactions between them. We show how variations of pore pressure and permeability can act in tandem to generate tremor-like patterns. The core feature of the model is that low-permeability plugs behave as elementary fault-valves. In a mechanism akin to erosive bursts and deposition events documented in porous media, valve permeability opens and closes in response to the local fluid pressure distribution. The rapid pressure release and/or mechanical fracturing associated with valve opening acts as the source of an LFE-like event. Valves interact constructively, leading to realistic, tremor-like patterns: cascades, synchronized bursts and migrations of activity along the channel, on both short and long time and space scales. Our model predicts that the input fluid flux is a key control on activity occurrence and behavior. Depending on its value, the subduction interface can be seismically quiescent or active, and seismicity can be strongly time-clustered, quasi-periodic or almost random in time. This model allows new interpretations of low frequency seismic activity in terms of effective fluid flux and fault-zone permeability.

Plain Language Summary

In between earthquakes, faults emit faint seismic signals, called tremor. Tremor is emitted as the fault channels fluid at very high pressures from deep sources towards the surface. As such processes can initiate earthquakes on the fault, they are the subject of intense scrutiny. In this work, we build a simple model to explore how fluids flow within a subduction fault and how it can be linked to tremor. In our model, the fault's conduits sometimes clog, because the fluid that circulates in it carries particles. As the fluid pushes through the plugs, it generates the seismic vibrations that we call tremor. If the fluid flux into the fault zone is just right, valves open and close in cycles, and the fault zone is permanently active. The patterns of model tremor it generates are very similar to the ones observed in real subduction zones: it migrates, and occurs in bursts of all sizes. However, if the flux is either too low or too high, the fault should remain silent. Our model

47 provides a framework to interpret seismic patterns as symptoms of how closed or open
48 a fault is to fluid circulation, and how strong the fluid flux is within it.

49 1 Introduction

50 The extensive records of weak low-frequency seismicity detected in many fault zones
51 over the last two decades provide powerful information on their dynamics. At interme-
52 diate depths (30–50 km) in several subduction zones, seismic activity takes the form of
53 tremor (*e.g.*, Obara, 2002; Kao et al., 2005; Payero et al., 2008; Husker et al., 2012). It
54 now seems clear that tremor is made of swarms of low-frequency earthquakes (LFEs) and
55 many authors use the term “tremor” as a synonym for LFE activity (*e.g.*, Shelly et al.,
56 2006; Song et al., 2009; Peacock et al., 2011; Audet & Kim, 2016; Frank et al., 2013, 2014;
57 Rogers, 2003; Bostock et al., 2012). These LFEs appear to be due to repetitive, impul-
58 sive and localized sources (Beroza & Ide, 2011; Frank et al., 2014). High rates of tremor
59 activity are associated with geodetically observed slow deformation episodes called slow-
60 slip events (SSEs) (*e.g.*, Kostoglodov et al., 2010; Frank, Radiguet, et al., 2015; Rous-
61 set et al., 2019; Delbridge et al., 2020). The correlation between LFE and SSE activity
62 gave rise to the term Episodic Tremor and Slip (ETS) (Rogers, 2003). In a more gen-
63 eral acceptance, transient seismic and geodetic signals that are both generated during
64 inter-seismic phases are often lumped together as *slow earthquake phenomena*. This new
65 class of seismic and geodetic observations holds great promise for improving our under-
66 standing of the seismo-tectonic cycle.

67 According to the framework that prevails today, slow earthquakes are the “slow”
68 analogs of “regular” earthquakes. Geodetic and seismic events are manifestations of the
69 same process of slow slip at transitional depths in fault zones, in between unstable stick-
70 slipping and stable sliding areas. The transitional part is made of slow sliding patches
71 that are responsible for geodetically detected SSEs and that contain many small embed-
72 ded frictionally unstable asperities generating LFEs. This framework explains why SSEs
73 and tremor activity are correlated in time and accounts for the S-wave dominated LFE
74 radiation pattern (*e.g.*, Shelly et al., 2007). Numerical models of these phenomena (*e.g.*,
75 Shibazaki & Iio, 2003; Liu & Rice, 2007; Ben-Zion, 2012; Ide, 2014; Luo & Liu, 2019; Bar-
76 bot, 2019) are often based on the rate-and-state friction (RSF) formalism developed for
77 “regular” earthquakes (Dieterich, 1992). This efficient modeling approach has two sig-
78 nificant shortcomings, however (van den Ende et al., 2018). First, all the dynamic vari-

79 ability of the interface and its surroundings is collapsed onto a planar boundary. Sec-
80 ond, it depends on empirical RSF parameters that are not inferred directly from pro-
81 cesses occurring within the fault zone. Thus, it is difficult to specify RSF parameters and
82 account for internal fault zone dynamics in a self-consistent manner. These shortcom-
83 ings may be limiting in the brittle-to-ductile transition zone of subduction faults because
84 the deformation proceeds over an increasingly wide off-fault region as depth increases
85 (Platt et al., 2018). Source dimensions for tremor and LFEs are typically inferred to be
86 a few hundred meters (*e.g.*, Chestler & Creager, 2017; Supino et al., 2020; Farge et al.,
87 2020; Sammis & Bostock, 2021), close to the thickness of the highly deformed perme-
88 able fault zone (*e.g.*, Angiboust et al., 2015). This calls for more complete models deal-
89 ing with processes in the interior of these zones.

90 There can be little doubt that pore fluid pressures and their variations play a key
91 role in fault zone dynamics (Saffer & Tobin, 2011). From a frictional perspective, high
92 pore pressures reduce the fault’s strength, bringing it closer to unstable sliding condi-
93 tions. Numerous geophysical and geological observations provide evidence for fluid sat-
94 uration in the subduction interface (Saffer & Tobin, 2011). It is widely believed that flu-
95 ids are supplied by metamorphic dehydration reactions in the downgoing plate (Manning,
96 1997; van Keken et al., 2011; Frezzotti & Ferrando, 2015; Hyndman et al., 2015). High
97 values of the V_p/V_s ratio and electrical conductivity indicate that pore fluids are at near-
98 lithostatic pressures (Peacock et al., 2011; Wannamaker et al., 2014; Audet & Kim, 2016,
99 among others), implying that the fault should be very weak. This explains why seismic
100 activity is highly sensitive to very small dynamic stress changes generated by passing tele-
101 seismic waves (Rubinstein et al., 2009) or tidal deformation (Rubinstein et al., 2008; Thomas
102 et al., 2009; Beeler et al., 2013; Royer et al., 2015).

103 There is mounting evidence for large variations of pore pressure on the time scale
104 of the slow-earthquake cycle. During slow slip events, seismic images, gravimetric stud-
105 ies as well as earthquake source mechanisms indicate that pore fluid pressure around the
106 subduction interface drops markedly (Tanaka et al., 2018; Warren-Smith et al., 2019; Gos-
107 selin et al., 2020). The migration of LFE activity is highly suggestive of pore-pressure
108 diffusion (Frank, Shapiro, et al., 2015; Cruz-Atienza et al., 2018). In exhumed subduc-
109 tion interfaces, the brittle-to-ductile transition seems to be intrinsically coupled to tran-
110 sient pore-pressure cycling and fluid transport (Angiboust et al., 2015; Platt et al., 2018;
111 Behr & Bürgmann, 2021). Episodes of fluid accumulation and drainage have been doc-

112 umented in shallow thrust faults and are recorded by metasomatic reactions in host rocks
113 (Angiboust et al., 2014; Saffer, 2015; Taetz et al., 2018). They have been attributed to
114 changes of permeable pathways due to tectonic deformation, precipitation/dissolution
115 or deposition/erosion processes (Giger et al., 2007; Jäger et al., 2017a; Williams et al.,
116 2019). In recent efforts to model ETS phenomena, the emphasis has shifted from a static
117 view of the impact of fluid pressure on RSF parameters to dynamic coupling between
118 time-dependent fluid pressure and slip (*e.g.* Bernaudin & Gueydan, 2018; Cruz-Atienza
119 et al., 2018; Zhu et al., 2020).

120 With the recognition that pore fluid pressures are high and variable in the subduc-
121 tion interface, it may not be appropriate to attribute seismic events to seismic slip only,
122 without considering other processes that are active in the fault zone. Indeed, there are
123 indications that the high fluid pressure transients that seem fundamental in shaping tremor
124 patterns may well generate seismic events on their own. Observations do not demonstrate
125 unambiguously that slow slip events and low-frequency seismic radiation are generated
126 by the same process. It is true that these two types of activity are frequently correlated
127 in time, but the sources of seismic and geodetic events are not systematically co-located
128 (*e.g.* Hirose et al., 2010; Kostoglodov et al., 2010). In addition, in some subduction zones,
129 LFEs do not conform to the moment-duration scaling of classical shear rupture (Bostock
130 et al., 2015; Farge et al., 2020) and do not follow the Gutenberg-Richter size distribu-
131 tion characteristic of regular earthquake activity (Sammis & Bostock, 2021). Finally, it
132 should be noted that the predominance of S-waves in the seismic signals does not nec-
133 essarily establish a shear rupture mechanism. S-wave dominated signals can arise from
134 volumetric sources with non-spherical geometries (*e.g.* Melnik et al., 2020) or from single-
135 force sources associated with fast local pressure variations (Shapiro et al., 2018).

136 Strong arguments in favor of a fluid-related origin for LFEs comes from laboratory
137 studies and volcano seismology. In volcanic systems, shear deformation and slip do not
138 play a predominant role, and yet tremors and LFEs (labelled as long-period earthquakes)
139 are generated in abundance. These events have been attributed to fluid related processes,
140 such as resonance in fluid-filled cracks (Chouet, 1996), strong flow in tortuous channels
141 (Julian, 1994; Benson et al., 2008) and rapid pressurization of volatiles (Melnik et al.,
142 2020; Wech et al., 2020), among others. In some cases, their sources are distributed over
143 large depth intervals of a few tens of kilometers and appear to migrate on time scales
144 of a few months or days (Shapiro et al., 2017). These migrations have been attributed

145 to pore-pressure transients associated with hydraulic connections between different parts
146 of a volcanic plumbing system.

147 A key piece of information on fault zone dynamics is the distribution of LFEs in
148 space and time. Here, we explore the potential of pressure transients in a permeable sub-
149 duction interface to generate this type of activity. The structure and hydraulics of such
150 a channel have rarely been studied in relation to seismicity (Segall & Rice, 1995; Aochi
151 et al., 2013; Cruz-Atienza et al., 2018; Zhu et al., 2020). We investigate the behavior of
152 plugs of low permeability in a heterogeneous channel. Pore pressure variations can arise
153 in two different ways. They may be generated by a time-dependent fluid supply, but there
154 are no known mechanisms for strongly fluctuating rates of metamorphic dehydration re-
155 actions. Alternatively, one can invoke changes of permeability, as in the classical fault-
156 valve model (Sibson, 1992). In this model, an increase of pore fluid pressure can gener-
157 ate an earthquake rupture, which opens up or reactivates fluid pathways leading to an
158 increase of permeability and a decrease of pore pressure. Clogging of pores due to pre-
159 cipitation from circulating fluids or to the trapping of particles then acts to decrease per-
160 meability, implying an increase of pore pressure and a new cycle of pressure change. The
161 permeability increase of the first phase of the cycle can be achieved in two different ways.
162 One is shear displacement in association with fault dilation (Im et al., 2019) — although
163 it appears that, in some cases, this may in fact act to decrease permeability in the slip
164 direction (Auradou et al., 2005). In the other mechanism, high pore pressures can in-
165 duce fracturing and/or unclogging of small pores and pore throats. In the laboratory,
166 impulsive flushing events similar to hydrofracturation have been observed in densely packed
167 permeable beds due to sudden particle detachment and called *erosive bursts* (Jäger et
168 al., 2017b; Bianchi, Thielmann, et al., 2018). These bursts are associated with sharp pore
169 pressure drops and recur as particles are redeposited. Coupled variations of permeabil-
170 ity and pore fluid pressure due to particle clogging and un-clogging have been observed
171 in many laboratory experiments (Candela et al., 2014; Bianchi, Thielmann, et al., 2018)
172 and have been surmised for fault zones (Manga et al., 2012). These phenomena have also
173 been documented in large-scale systems. In some oil fields, for example, long-term records
174 show that the bulk permeability of the reservoir oscillates in relation to cycles of depo-
175 sition and removal of asphalt aggregates (Sahimi et al., 1999). We therefore argue that
176 these mechanisms are bound to be active in faults, where fluid circulates under high pres-
177 sure gradients, carrying precipitates and cataclastic fines that can be deposited and dis-

178 lodged. In this framework, fast local pore pressure variations (Shapiro et al., 2018) and/or
179 rupture of brittle heterogeneities due to increased fluid pressure (Fagereng & Sibson, 2010;
180 Kotowski & Behr, 2019) may be responsible for LFEs. We expect that the unclogging
181 or rupture of an individual plug induces pore pressure variations that propagate to neigh-
182 boring plugs and generates swarms of similar events.

183 Through the rapid pressure drop that occurs, the unclogging of an individual plug
184 is responsible for a local impulsive force on the solid matrix which generates elastic waves
185 with both compressional and shear polarisation components. Shapiro et al. (2018) have
186 shown that the radiation pattern of such a single-force source is consistent with seismic
187 observations in subduction zones due to the constrained source-receiver geometry, and
188 that it can account for the weak dependency of LFEs source duration on their magni-
189 tude, as observed in Cascadia and Guerrero (Bostock et al., 2015; Farge et al., 2020).
190 Here, we build on this study to evaluate the seismic patterns that can be generated by
191 a series of interacting plugs in a permeable channel. The clogging/unclogging mechanism
192 does not involve shear rupture explicitly and we will discuss *a posteriori* how it should
193 reflect on slip and its sensitivity at high fluid pressures.

194 The paper is organized as follows. We first describe the main characteristics of tremor/LFE
195 activity in subduction zones that must be accounted for by any physical model. We then
196 develop a quantitative model, starting with the basic processes that are involved in the
197 clogging and unclogging of a plug and the corresponding equations. Results are described
198 for a range of fluid input rates and a regime diagram is obtained and justified by sim-
199 ple theoretical arguments. The activity patterns that are generated are compared to ob-
200 served ones, and conditions that must be met for a successful model are specified.

201 **2 Tremor patterns observed in subduction zones**

202 We summarize the most salient features of tremor activity in subduction zones. The
203 Guerrero, Mexico, area is particularly appropriate to study LFE activity because it stretches
204 over a distance of more than 100 km along-dip, owing to a low dipping angle. A very large
205 catalog of events is available there (Frank et al., 2014).

206 **2.1 Spatial Segmentation**

207 At a local scale, the fact that LFEs occur in families of recurring events with sim-
 208 ilar waveforms indicates that they are due to the repeated activation of individual sub-
 209 kilometric scale sources (Chestler & Creager, 2017). In subduction zones, activity is of-
 210 ten segmented in the along-strike direction and concentrated in patches (Brudzinski &
 211 Allen, 2007; Poiata et al., 2020). In Shikoku, Japan, and elsewhere, events delineate pluri-
 212 kilometric striations that are aligned with the subduction direction and that have been
 213 linked to the subduction of seamount chains (Ide, 2010). Shorter striations have also been
 214 observed in the Olympic Peninsula, Cascadia (Ghosh et al., 2010). In Guerrero, Mex-
 215 ico, activity is split into two well-separated patches along the dip of the interface (Payero
 216 et al., 2008; Kostoglodov et al., 2010; Husker et al., 2012; Frank et al., 2014) (Figure 1).

217 **2.2 Tremor Migration**

218 Bursts of tremor activity migrate along the subduction interface on time scales of
 219 seconds to weeks, and spatial scales from a few to a few hundreds of kilometers. Short-
 220 lived, short-distance migrations are often restricted to an individual tremor patch and
 221 proceed at rates in the 10–100 km/hr range (on time scales T of minutes to hours over
 222 length scales L of tens of kilometers) (Ghosh et al., 2010; Cruz-Atienza et al., 2018). In
 223 marked contrast, long-lived, long-distance migrations that connect spatially separated
 224 patches are slower, with velocities in the ~ 1 –10 km/day range ($T \sim$ weeks to months,
 225 $L \sim 100$ km) (Frank, Shapiro, et al., 2015; Kao et al., 2009). Such systematics between
 226 time and space scales are reminiscent of a diffusion process, such that $L^2 \propto T$ and there-
 227 fore $V \sim L/T \propto 1/L$.

228 In Japan and Cascadia, the largest transients of activity ($T \sim$ weeks, $L \sim 100$ km)
 229 propagate along-strike in a belt lying in a 30 to 60 km depth range. It has been suggested
 230 that this depth interval delineates a zone rheologically conducive to slow-slip (Ide, 2012;
 231 Gomberg et al., 2016). In Guerrero, Mexico, the active segment of the subducted slab
 232 is dipping at a low angle and stretches over as much as 200 km in a 40–50 km depth range
 233 (Pardo & Suárez, 1995; Kim et al., 2010). Due to this particularly favorable configura-
 234 tion, it is clear that the largest migrations systematically originate from the deep end
 235 of the fault and slowly propagate updip (Frank, Shapiro, et al., 2015).

236 Short-lived migrations ($T \sim$ minutes to hours) often proceed along striations (Ide,
 237 2012), sometimes in rapid sweeps (Ghosh et al., 2010). Remarkably, these rapid migra-
 238 tions sometimes occur within larger scale migrations, propagating in the reverse direc-
 239 tion with respect to the large scale propagation. This has been observed in the along-
 240 dip direction in Guerrero, Mexico (Frank et al., 2014; Cruz-Atienza et al., 2018) and in
 241 the along-strike direction in Cascadia (Houston et al., 2011).

242 **2.3 Temporal Patterns**

243 Tremor activity occurs during episodic *bursts* of events, which are composed of shorter
 244 bursts. Temporal clustering of activity covers time scales of years for the largest bursts
 245 — often correlated with slow-slip events and long-distance migrations — to days or even
 246 minutes for smaller bursts (Wech & Creager, 2011; Idehara et al., 2014) — called *cas-*
 247 *cades*. The imbrication of small bursts within larger ones makes it difficult to determine
 248 a characteristic duration. The shortest bursts proceed over a few minutes, whereas the
 249 largest ones can persist for days or even months (*e.g.* Ghosh et al., 2010; Frank et al.,
 250 2014; Shelly, 2015) As a temporal point process, this activity is clustered and does not
 251 conform to simple memoryless realizations. It has been analyzed using non-Poissonian
 252 statistics (Frank et al., 2016; Beaucé et al., 2019).

253 The recurrence interval between tremor bursts decreases with increasing depth along
 254 the subduction interface (Wech & Creager, 2011). In addition, the time-clustering of ac-
 255 tivity can change from one tremor patch to its neighbors (Idehara et al., 2014; Frank et
 256 al., 2016), and can evolve as a function of time (Frank et al., 2016). This has been in-
 257 terpreted as evidence for variations of fault zone conditions in both space and time (Frank
 258 et al., 2016).

259 **2.4 Tremor activity Patterns in Guerrero, Mexico**

260 Figure 1 shows the spatio-temporal patterns of LFE activity in the Guerrero tremor
 261 subduction segment, which is concentrated in two separate patches. One, called the *sweet*
 262 *spot*, is located at the downdip end of the segment and has a high rate of activity with
 263 a high level of time clustering (Frank et al., 2016). The other patch, called *transient zone*
 264 (Frank et al., 2014), is further updip and on the whole more quiescent. It hosts episodic
 265 bursts of activity with a recurrence interval of about 3 months (Frank et al., 2014), which

266 coincide with short-term SSEs recurrence (Frank, Radiguet, et al., 2015). One of these
 267 episodes is displayed in Figure 1c. Activity migrates away from the sweet spot over a
 268 hundred kilometers updip. These episodes are built from many shorter bursts that may
 269 drift in both updip and downdip directions (Figure 1d).

270 The Guerrero subduction zone is also known for its very large SSEs that recur at
 271 intervals of about 4 years (*e.g.*, Radiguet et al., 2012) and are accompanied by intense
 272 tremor activity in both the transient zone and sweet spot. These events persist for sev-
 273 eral months with slow slip occurring mostly updip of the tremor-generating segment. The
 274 unstable slipping segment of the fault encroaches upon the transient zone over a small
 275 distance, where short-term SSEs are concentrated. During both short and long-term SSEs,
 276 tremor activity in the transient zone switches from episodic to highly clustered (Frank
 277 et al., 2016).

278 The most striking feature of the Guerrero subduction zone is perhaps that the patch
 279 with maximum tremor activity (the sweet spot) is separated from the patch with largest
 280 amount of slow slip (the transient zone) by more than 20 km.

281 **3 Model design**

282 In this section, we develop a model for time changes of permeability in a hetero-
 283 geneous porous channel that runs along the dip of the subduction interface. Based on
 284 physical arguments, we associate sudden changes of fluid pressure with the generation
 285 of LFE-like, elementary tremor events.

286 **3.1 Diffusion equation for pore pressure variations**

287 Pore connectivity is modeled at a macroscopic level with a permeability that varies
 288 along the channel $k(x)$. The fluid mass flux per unit cross-sectional area, $q(x, t)$, depends
 289 on pore pressure $P(x, t)$ according to Darcy's law:

$$q(x, t) = -\frac{k(x)\rho}{\eta} \left(\frac{\partial P}{\partial x}(x, t) + \rho g \sin \alpha \right) \quad (1)$$

290 where ρ is the fluid's density, η is the fluid viscosity, α is the dip angle of the subduc-
 291 tion interface and g the acceleration of gravity. Mass balance dictates that :

$$-\frac{\partial q}{\partial x}(x, t) = \frac{\partial(\rho\phi)}{\partial t}. \quad (2)$$

292 where ϕ is porosity. We consider that pores are saturated with fluid and neglect tem-
 293 perature changes, which is justified by the small thickness of the channel and the small
 294 flow rates that are expected. Changes of the local mass of fluid can be due to the fluid
 295 compressibility as well as expanding/contracting pores. This is accounted for by an ef-
 296 fective rock-fluid compressibility β , such that:

$$\begin{aligned} \beta &= \frac{1}{\rho\phi} \frac{\partial(\rho\phi)}{\partial P} \\ \implies -\frac{\partial q}{\partial x}(x, t) &= \rho\phi\beta \frac{\partial P}{\partial t}(x, t). \end{aligned} \quad (3)$$

297 For simplicity, we neglect density variations in the body force term in equation (1), be-
 298 cause they are due to changes of flow-induced dynamic pressure, which account for a frac-
 299 tion of the total pressure, and because they are not expected to play a significant role.
 300 We split pressure into a hydrostatic component and a time-dependent dynamic compo-
 301 nent due to flow noted p . Neglecting changes of fluid viscosity, we obtain the following
 302 equation for the dynamic pore pressure component:

$$\frac{\partial p}{\partial t} = \frac{1}{\phi\beta\eta} \frac{\partial}{\partial x} \left(k(x) \frac{\partial p}{\partial x} \right). \quad (4)$$

303 We assume that permeability is piecewise constant. Thus, in each segment, the dynamic
 304 pressure obeys a diffusion equation:

$$\frac{\partial p}{\partial t} = \frac{k}{\phi\beta\eta} \frac{\partial^2 p}{\partial x^2}, \quad (5)$$

305 with diffusivity $D = k/(\phi\beta\eta)$. Along the modeled channel, D is therefore piecewise con-
 306 stant.

307 Figure 2 shows how we simplify the subduction fluid transport system conceptually
 308 and how it is represented as a permeable channel with plugs of low permeability.

309 **3.2 A mechanism for large changes of permeability**

310 Our model has two main ingredients, which deal with the nature of the fluid phase
 311 and its interaction with the network of pores in the permeable medium. Subduction zone

312 metamorphic fluids are complex silica-saturated mixtures (Manning, 1997; Frezzotti &
 313 Ferrando, 2015). As they rise, they undergo decompression and, due to the pressure de-
 314 pendence of solubility, are expected to precipitate either amorphous silica (silica gel) or
 315 quartz nanoparticles as well as other mineral phases. Depending on the availability of
 316 nucleation sites, precipitation may proceed along the pore walls or in the bulk fluid. High
 317 temperature decompression experiments on silica-rich fluids demonstrate that the sil-
 318 ica nanoparticles that are generated are readily mobilized by fluid flow (Amagai et al.,
 319 2019; Okamoto, 2019). Small clasts generated by grinding motion along the fault are also
 320 likely involved (*e.g.* Han & Hirose, 2012). As in other deep permeable systems, the in-
 321 terstitial fluid phase is thus expected to carry fines and colloids (Manga et al., 2012).

322 Natural permeable media are made of pores and fractures connected by thinner throats
 323 which have a strong effect on permeability because they determine the effectiveness of
 324 fluid pathways through a volume. Capture and/or precipitation of even small amounts
 325 of solids within throats can decrease permeability by orders of magnitude (McDowell-
 326 Boyer et al., 1986). Calculations by Beckingham (2017) show that, depending on whether
 327 pore throats are open or clogged, permeability can change by more than a factor of ten.
 328 These changes are associated with negligible porosity variations because pore throats oc-
 329 cupy a very small fraction of the total interstitial volume. Thus, the transition between
 330 high and low permeability states is very abrupt, as plugging thin throats between larger
 331 pores involves very small amounts of sediment. Even if they are much smaller than a pore
 332 throat, fines and colloids can get stuck through a jamming process at rates that depend
 333 on their concentration (Agbangla et al., 2014; Civan, 2016; Delouche et al., 2020). Clog-
 334 ging can be achieved in a matter of minutes (Manga et al., 2012; Candela et al., 2014;
 335 Delouche et al., 2020).

336 Material that gets stuck in pore throats may be remobilized by the flow (Civan,
 337 2016; Kudrolli & Clotet, 2016; Jäger et al., 2017b; Bianchi, Wittel, et al., 2018; Ama-
 338 gai et al., 2019). Flow exerts a shear stress on the encasing solid, noted σ , that scales
 339 with the average pathway diameter d and the pressure gradient $\partial p/\partial x$. If this stress ex-
 340 ceeds a certain threshold value noted σ_c , infillings are eroded and transported by the fluid
 341 (Civan, 2016). This may occur in catastrophic fashion in events called *erosive bursts* (Bianchi,
 342 Thielmann, et al., 2018). Such bursts may be composites of several smaller unclogging
 343 events (Bianchi, Thielmann, et al., 2018). The process has been studied in detail by Kudrolli
 344 and Clotet (2016). In these experiments, a pure fluid is fed into a polydisperse particle

345 bed. When the flow rate exceeds a threshold value, such that the flow-induced stress ex-
 346 ceeds the threshold value σ_c , the flow dislodges and mobilizes the smaller particles. In-
 347 terestingly, the authors observe small precursor events when the flow rate is just below
 348 critical and a large avalanche when the flow rate reaches the critical value. This is due
 349 to the collective behavior of particles, as some of them are not able to move if they are
 350 in contact with stronger ones that resist detachment. Thus, these particles are only dis-
 351 lodged when their neighbors are, which triggers an erosion avalanche. As long as $\sigma >$
 352 σ_c , particles cannot remain stuck and incipient coatings of pore walls are eroded, ensur-
 353 ing that fluid pathways stay fully open. Clogging thus requires that $\sigma \leq \sigma_c$. The al-
 354 ternation between erosion and deposition phases leaves little opportunity for particles
 355 and colloids to become permanently cemented to one another and to pore walls. Clog-
 356 ging and unclogging events therefore proceed rapidly, and can be considered as instan-
 357 taneous at the time-scales of LFE activity.

358 As shown above, these processes have been observed in laboratory experiments on
 359 different types of porous media. Owing to their local character, they have not been doc-
 360 umented in the field but, on a large scale, simultaneous variations of permeability and
 361 fluid pressure in faults (Manga et al., 2012) and oil reservoirs (Sahimi et al., 1999) have
 362 been attributed to erosion/deposition processes.

363 **3.3 A valve mechanism**

We model clogging and unclogging events as follows. Within a long permeable chan-
 nel, let us consider a clogged plug where fluid pathways are narrow and contorted be-
 cause they wrap around captured particles (closed state). In this case, the average path-
 way aperture d and permeability k are both small, with values d_{lo} and k_{lo} , respectively.
 If and when δp , the pressure difference across the plug, reaches a critical value noted δp_c^{break}
 such that the shear stress on the plug particles σ is equal to critical value σ_c , unclogging
 occurs. As a consequence, pore throats are no longer obstructed by particles and per-
 meability increases to $k_{hi} \gg k_{lo}$, in association with an increase of d from d_{lo} to d_{hi}
 (open state). This drives a lowering of pore pressures upstream of the plug, and an in-
 crease downstream. This pressure transient is very rapid initially, and then decelerates,
 in a diffusion-induced flow transient. A key point is that σ initially increases. Just be-

fore unclogging, it is equal to:

$$\sigma_- = \lambda_1 d_{lo} \frac{\delta p_c^{break}}{w} = \sigma_c \quad (6)$$

where λ_1 is a coefficient and w is the plug length along the channel. As plug particles are set in motion by the hydraulic stress and can no longer adhere to throat walls, the effective pore aperture d increases from d_{lo} to d_{hi} . In the permeable medium, variables are defined as spatial averages over a representative volume, and this is true for porosity and pressure in particular. As the solid particles get flushed out of the thin throats that connect larger pores, porosity is barely affected and, by the same token, so is the fluid pressure, because it is the average pressure for a group of pores and pore throats, with the larger pores acting to buffer local pressure variations at the pore throat scale. Thus, the pressure differential across the pore is maintained at δp_c^{break} as the plug opens up (Jäger et al., 2018). Therefore, σ increases to σ_+ :

$$\sigma_+ = \lambda_1 d_{hi} \frac{\delta p_c^{break}}{w} \quad (7)$$

which is larger than σ_c . σ then decreases gradually as the pressure distribution adapts to the new permeability structure. The plug remains unobstructed as long as σ is larger than σ_c . Clogging occurs when $\sigma = \sigma_c$ such that δp reaches a second threshold value of pressure differential noted δp_c^{clog} :

$$\lambda_1 d_{hi} \frac{\delta p_c^{clog}}{w} = \sigma_c. \quad (8)$$

364 Clogging causes the permeability to drop back to k_{lo} , driving an increase in δp until un-
 365 clogging occurs again, thus starting a new cycle. This is a valve behavior akin to the toggle-
 366 switch of Miller and Nur (2000).

This framework allows an internally consistent mechanism for a succession of unclogging and clogging events. One requirement is that $\delta p_c^{clog} < \delta p_c^{break}$. From equations (6) and (8), we deduce that:

$$\frac{\delta p_c^{break}}{\delta p_c^{clog}} = \frac{d_{hi}}{d_{lo}}. \quad (9)$$

Thus, the fact that $d_{hi} > d_{lo}$ automatically ensures that $\delta p_c^{clog} < \delta p_c^{break}$. One can go one step further using relationships between the average width of fluid pathways d and permeability k . A common parameterization is the Kozeny-Carman equation (Civan, 2016):

$$k = \lambda_2 \frac{\phi d^2}{\tau} \quad (10)$$

where λ_2 is a coefficient that depends on the nature of the fluid pathways, *e.g.* tubes or slits, ϕ is porosity and τ is tortuosity. We may neglect variations of porosity, as discussed above, and obtain:

$$\frac{k_{hi}}{k_{lo}} = \frac{\tau_{closed}}{\tau_{open}} \left(\frac{d_{hi}}{d_{lo}} \right)^2 \quad (11)$$

where τ_{open} and τ_{closed} are the tortuosities of the open and closed plugs. Fluid pathways are more irregular and contorted in a closed plug than in an open one, implying that $\tau_{closed} > \tau_{open}$. Using equation (9), one deduces that:

$$\frac{\delta p_c^{break}}{\delta p_c^{clog}} = \frac{d_{hi}}{d_{lo}} = \left(\frac{\tau_{open}}{\tau_{closed}} \frac{k_{hi}}{k_{lo}} \right)^{1/2}. \quad (12)$$

367 For the small porosity media that are involved here, tortuosity ratio $\tau_{open}/\tau_{closed}$ is likely
 368 to be in a 1/2–1/4 range (Ghanbarian et al., 2013). As stated above, permeability changes
 369 induced by clogging are typically larger than one order of magnitude (McDowell-Boyer
 370 et al., 1986; Beckingham, 2017), which guarantees that $\delta p_c^{break} > \delta p_c^{clog}$. For the sake
 371 of further discussion in section 4.1, it should be noted that such values of the tortuos-
 372 ity ratio imply that $k_{hi}/k_{lo} > \delta p_c^{break} / \delta p_c^{clog}$.

373 Figure 3 summarizes how these processes are implemented in our model. In valve
 374 segments of the channel, permeability responds to the local fluid pressure gradient ac-
 375 cording to the hysteretical cycle shown in Figure 3b.

376 **3.4 Valves as elementary seismic sources**

377 At each valve-opening, the fluid pressure gradient that has built up is suddenly dis-
 378 sipated in a rapid, localized diffusion transient. Variations of fluid pressure on each sides
 379 of the valve are responsible for temporal changes of the force applied to the solid walls
 380 of the permeable medium, which are able to generate seismic waves, as shown by Shapiro
 381 et al. (2018). Other seismic processes may also be triggered. For example, high fluid pres-
 382 sures may induce the seismic rupture of critically-stressed brittle heterogeneities (*e.g.*
 383 Taetz et al., 2018), and an impulsive turbulent flow transient may generate resonance
 384 in tortuous pathways (Benson et al., 2008). We thus consider that as they open, valves
 385 act as elementary seismic sources and generate LFE-like events.

386 **3.5 Permeability values, boundary and initial conditions**

387 We will consider heterogeneous permeable channels with a series of randomly dis-
 388 tributed valves. This corresponds to one type of random permeability distribution in-

389 volving segments where fluid pathways are more contorted than elsewhere. The behav-
 390 ior of any valve depends on what happens in neighboring ones, so that pressures and flow
 391 rates must be calculated everywhere in the channel.

392 Metamorphic reactions occur at different depths depending on the dehydrating ma-
 393 terial, which includes sediments, serpentinized mantle and hydrated oceanic crust. Fo-
 394 cussing on the oceanic crust, dehydration proceeds at depths exceeding the tremor source
 395 region (van Keken et al., 2011) and most of this fluid gets trapped in the high perme-
 396 ability subduction interface, sealed from above by the low permeability mantle wedge
 397 (Hyndman et al., 2015). Thus, the channel is fed at its base by an input fluid flux q_{in} .
 398 For a first instalment of the model, it is instructive to maintain this flux at a constant
 399 value. At its updip end, the channel is connected to a high permeability region which
 400 may be either continental crust (Hyndman et al., 2015) or an accretionary prism. In this
 401 case, pore pressures are close to hydrostatic values, such that the dynamic pressure is
 402 zero ($p_{out} = 0$).

403 For simplicity, we assume that in the channel outside of valves, permeability remains
 404 constant at the value k_{hi} and that valves have the same length w in the x direction. Phys-
 405 ical variables are made dimensionless using scales described in Table 1. Values adopted
 406 for the various valve parameters are listed in Table 2. Boundary and initial conditions
 407 are illustrated in Figure 2b,c. We start all simulations with an equilibrium profile cor-
 408 responding to a low value of flux $q_{in}(t < 0) = 0.09$, such that all valves are closed with
 409 permeability k_{lo} . At $t = 0$, q_{in} is changed to its value for the run.

410 **3.6 Numerical implementation**

411 Solutions for the pressure field in time were obtained using a Crank-Nicholson fi-
 412 nite difference scheme, with a first order, centered discretization of the second order space
 413 derivatives of pressure (Press et al., 2007, section 20.2, pp 1045–1048). The scheme is
 414 implicit, centered and second order in time, and is unconditionally stable. Valves are ei-
 415 ther in an open or a closed state, as described in section 3.3 above, and when the pres-
 416 sure differential at their boundary reaches a threshold value, their open/closed state is
 417 changed accordingly. Each opening of a valve is counted as a seismic event.

418 In all simulations presented here, distance along the fault is discretized in dimen-
 419 sionless increments $\delta x = 0.002$ and the dimensionless time step is set to $\delta t = h^2/2D =$

420 2.10^{-6} . In order to test the accuracy of the numerical implementation, we calculated an-
 421 alytically the pressure variations that follow an unclogging event in and around an iso-
 422 lated valve, and compared them to numerical solutions. Convergence was then tested
 423 by lowering both space and time steps, and we verified that accuracy improved with in-
 424 creasingly fine discretization. Section S1 in the supplementary information file describes
 425 the methodology and results of the convergence and accuracy tests.

426 **4 Time-dependent valve behavior**

427 We first use a single valve to show how the activity regime is controlled by the in-
 428 put flux and how an intermittent valving behavior requires certain conditions (Figures 4
 429 and 5). We then illustrate how neighboring valves interact using 3 valves (Figure 6).

430 **4.1 A single valve**

As mentioned above, the valve mechanism depends on five control parameters, four
 that describe valve behavior, $k_{lo}, k_{hi}, \delta p_c^{clog}, \delta p_c^{break}$, and the input flux q_{in} . We identify
 the conditions under which a steady state solution exists, with a valve that is either open
 or closed. In steady state, the flux through each valve is equal to the input flux:

$$q_{in} = \frac{\rho}{\eta} k_v \frac{\delta p}{w} \quad (13)$$

431 where k_v , the valve permeability, may be equal to k_{lo} or k_{hi} , and where δp is again the
 432 pressure difference across the valve. If $\delta p \leq \delta p_c^{break}$ in a closed valve, stuck particles
 433 cannot be removed and the valve remains closed. Conversely, if $\delta p \geq \delta p_c^{clog}$ in an open
 434 valve, particles cannot be captured and the valve remains open. These conditions can
 435 be expressed as follows, using q_{in} , the flux a valve is subjected to:

$$q_{in} \leq \frac{\rho}{\eta} k_{lo} \frac{\delta p_c^{break}}{w} \quad (= q_c^{break}) \quad \text{for a closed valve,} \quad (14)$$

$$q_{in} \geq \frac{\rho}{\eta} k_{hi} \frac{\delta p_c^{clog}}{w} \quad (= q_c^{clog}) \quad \text{for an open valve.} \quad (15)$$

436 It introduces two threshold values for the fluid flux, noted q_c^{break} and q_c^{clog} , respectively.
 437 The two conditions in (14) and (15) define domains in $(\delta p, k_v)$ space corresponding to
 438 a closed valve or an open one in steady-state (Figure 4). Table 2 lists values for the pres-
 439 sure thresholds and permeabilities, as well as for the flux thresholds used for the valve
 440 set up of this study.

441 Depending on the four valve parameters, there are two different types of solutions.
 442 Let us first consider *enhanced permeability variations*, such that $k_{hi}/k_{lo} > \delta p_c^{break}/\delta p_c^{clog}$.
 443 In this case, $q_c^{break} < q_c^{clog}$ and there are no steady-state solutions in the $q_c^{break} - q_c^{clog}$
 444 interval (Figure 4a). In this interval, the input flux being higher than q_c^{break} , a closed
 445 valve will open, and in turn, the input flux being lower than q_c^{clog} , the now open valve
 446 will eventually close. Therefore, the valve must constantly switch between open and closed
 447 states in order to handle the imposed input flux. This oscillatory regime is illustrated
 448 schematically in Figure 3.

449 In the other case, called *enhanced pressure variations*, $k_{hi}/k_{lo} < \delta p_c^{break}/\delta p_c^{clog}$.
 450 This implies that $q_c^{break} > q_c^{clog}$ and the domains for a closed valve and an open one
 451 overlap in the $q_c^{clog} - q_c^{break}$ interval. In this interval, steady-state solutions are possi-
 452 ble for the two valve states (Figure 4b). Which one is selected depends on the initial state
 453 of the valve. In this case, therefore, there is no valve activity.

This analysis leads to the following instability criterion for the valve:

$$q_c^{break} < q_c^{clog} \iff \frac{\delta p_c^{break}}{\delta p_c^{clog}} < \frac{k_{hi}}{k_{lo}}. \quad (16)$$

454 In a geological context, the tortuosity ratio is small and lies in a restricted range
 455 (see section 3.3, last paragraph) and in contrast, permeability is expected to change by
 456 one order of magnitude (McDowell-Boyer et al., 1986; Miller & Nur, 2000; Beckingham,
 457 2017). Using the argument developed for equation (12), where we derive the threshold
 458 δp ratio as a function of the permeability ratio, we conclude that the instability crite-
 459 rion should be met in geological porous media where erosion/deposition processes are
 460 active. Such permeable systems have been reported to self-organize around these crit-
 461 ical erosion/deposition conditions (Kudrolli & Clotet, 2016).

462 Figure 5 shows results for three values of input flux, corresponding to steady-state
 463 with a closed valve ($q_{red} < q_c^{break}$), a permanent intermittent regime with an unstable
 464 valve that switches between closed and open states ($q_c^{break} < q_{yellow} < q_c^{clog}$), and fi-
 465 nally steady-state with an open valve ($q_{green} > q_c^{clog}$). In the first case, q_{red} is too low
 466 to allow the valve to open at all, no activity is recorded. In the last case, q_{green} is so high
 467 that after an initial transient during which the valve opens and closes to accomodate the
 468 incoming flux, the valve ends up open, submitted to a flux that does not allow it to close.
 469 In the intermediate case, q_{yellow} is both high and low enough to allow the valve to open

470 when closed, and to close when open. After an initial transient, the valve activates in
 471 a very regular fashion.

472 4.2 Valve-valve interaction via pore pressure diffusion

473 A simulation with three valves shows how the opening of a valve may trigger the
 474 opening of adjacent valves (Figure 6). In this simulation, the middle valve v_2 opens up,
 475 driving a sudden permeability increase. As a consequence, the δp that has built up across
 476 the valve re-equilibrates through diffusion, decreasing pressure updip of the valve and
 477 increasing it downdip of it. This induces an increase of pore pressure gradient through
 478 the two adjacent valves, eventually bringing them closer to unclogging conditions. We
 479 note that such constructive interaction causes migrations of activity in cascades, in both
 480 up- and downdip directions (see section 5.2).

481 5 Emergence of collective patterns for a large number of interacting 482 valves

483 We now explore systematically a model involving 29 randomly distributed valves
 484 (Figure 2b). With such a valve density, there are patches with valves that are closer to-
 485 gether than on average, thus strengthening interactions that may generate spatial pat-
 486 terns of activity. All the valves are identical. We take the same permeability and thresh-
 487 old δp as specified in Table 2, implying that the threshold flux q_c^{clog} and q_c^{break} are the
 488 same as in previous sections (equations (14) and (15)). In fault-zones, we expect that
 489 conditions for enhanced permeability variations are met (see section 4.1, equation (16)
 490 in particular) and hence focus on the case in which $q_c^{break} < q_c^{clog}$.

491 5.1 Characteristics of valve activity in the permanent regime

Model outputs for runs with low ($q_{in} = 0.16$) and high ($q_{in} = 0.81$) input fluxes
 are shown in Figure 7 and Figure 8. At any given time, a simple diagnosis of the state
 of the system is given by the bulk equivalent permeability k_{eq} . Considering that chan-
 nel segments and valves constitute hydraulic resistors in series, k_{eq} can be written:

$$k_{eq} = \frac{L}{(L - N_{cl}w)/k_{hi} + N_{cl}w/k_{lo}} \quad (17)$$

492 where N_{cl} is the number of closed valves, and L is the channel length. For any of the three
 493 regimes, activity rate, k_{eq} and the pressure difference across the channel Δp reach a dy-

494 namic equilibrium around a stable value after an initial transient, thus defining a per-
 495 manent regime (Figures 7 and 8).

496 The analysis of section 4.1 for a single valve does not depend on the presence of
 497 neighboring valves, and holds for a channel with multiple valves. The three permanent
 498 regimes for the 29-valve system are illustrated in Figure 9: one with intermittent valve
 499 activity and two quiescent ones. Valve activity occurs when the flux is within the q_c^{break} -
 500 q_c^{clog} range. Within that range, we find that the activity style changes as a function of
 501 the input flux (Figure 10). Near the two ends of that range, activity is episodic or time-
 502 clustered (see also Figures 7 and 8), and almost periodic for the highest values of q_{in} (as
 503 shown later on, Figures 11 and 12). For intermediate flux values in the middle of the flux
 504 range, activity is random and Poisson-like, with a seemingly constant overall rate. This
 505 result holds for the valve distribution displayed in Figure 2b, and for 30 other random
 506 distributions of 29-valves with the same characteristics (parameters in Table 2). This will
 507 be discussed in more detail below.

508 Here, the catalog of synthetic events in space and time is analyzed in the form of
 509 a two-dimensional point process (t_i, x_i) illustrated in time-location graphs (Figures 7d,
 510 8d, 11 and 12c). To characterize the level of time clustering (episodicity) of synthetic events,
 511 we follow Frank et al. (2016) and use statistical tools for the analysis of temporal point
 512 processes (Lowen & Teich, 2005). The main idea is to compute the spectrum of the event
 513 count signal autocorrelations, and to estimate the power-law exponent γ of its high fre-
 514 quency decay. A flat spectrum with $\gamma \approx 0$ corresponds to a nearly Poissonian process,
 515 whereas γ values that are significantly larger than 0 are indicative of time clustering.

516 As shown in Figure 10a, activity is Poisson-like for intermediate values of the in-
 517 put flux, such that $\gamma \approx 0$. For values of the input flux at both ends of the q_c^{break} - q_c^{clog}
 518 range, activity is characterized by high values of γ , indicating time clustering. Time-series
 519 of the activity rate reveal episodic bursts (Figure 10b). When the input flux is low (close
 520 to q_c^{break}), activity appears to be close to scale-invariant clustering with a burst recur-
 521 rence rate that depends on the time scale of the measurement. For higher values of the
 522 input flux close to the upper bound q_c^{clog} , activity proceeds in quasi periodic bursts, with
 523 periods that seem to increase with increasing q_{in} .

524 As the input flux is increased from the lower threshold value q_c^{break} to the upper
 525 one q_c^{clog} , the channel is increasingly open on average. This may be assessed from the

526 average proportion of open valves which increases as a function of q_{in} (Figure 10c). The
 527 proportion of open valves dictates the evolution of k_{eq} , the bulk permeability of the chan-
 528 nel as a whole. For a low input flux ($q_{in} \approx q_c^{break}$), valves are closed most of the time
 529 and activity bursts are linked to phases of opening. In the other extreme, for $q_{in} \approx q_c^{clog}$,
 530 permeability is open most of the time. Swarms of events then occur mainly when the
 531 channel shuts down temporarily and valve pressures increase until unclogging conditions
 532 are met.

533 5.2 Cascades and migrations of synthetic activity

534 In the random distribution of valves that was adopted here, there are three patches
 535 where valves are closer to one another than elsewhere, around $x = 0.1$, $x = 0.5$ and
 536 $x = 0.85$ (Figure 11). In patches of closely packed valves, the constructive interaction
 537 we describe in Figure 6 is responsible for the occurrence of rapid cascades of events. These
 538 migrations may proceed in both downdip and updip directions, apparently with simi-
 539 lar propagation rates. Obviously, the interaction is weaker and slower for valves that are
 540 separated by larger distances. Thus, the rates of cascading and event migration depend
 541 on the valve distribution.

542 Figure 11 depicts typical synthetic activity patterns. The input flux is close to the
 543 upper threshold q_c^{clog} , $q_{in} = 0.93$. Figure 11c shows nearly deterministic patterns of col-
 544 lective behavior at both short time and spatial scales. Cascades of triggering/opening
 545 events are associated with migrations of activity that show up as oblique dot alignments
 546 in the time-location chart. For this run as well as others, the migration rate of cascades
 547 takes similar dimensionless values $v \approx 4 - 6 \times 10^2$, with faster velocities (up to $v \approx$
 548 2×10^3 in dimensionless units) in the densest valve patches.

549 The fastest migrations appear to start and stop at the edges of densely populated
 550 patches. In such a patch, valves are close to one another and act in tandem. This shows
 551 up as an activity cluster which hosts its own internal migration patterns. Larger scale
 552 valve patches are activated sequentially in the updip direction. These large scale migra-
 553 tions proceed at slower rates than those within a single valve patch (Figure 11b). These
 554 large-scale migrations propagate at $V \approx 22 \times 10^2$ in dimensionless values from one
 555 end of the channel to the other.

5.3 Large-scale valving behavior

Segments of the channel containing a number of closely-spaced valves act like macroscopic valves, such that their equivalent permeability varies in hysteretical cycles with the total pressure difference across them. This behavior is achieved over a large range of scales, from small sets of a few valves to large valve patches and eventually for the full channel. Figure 12 depicts the cycles of equivalent permeability k_{eq} and total pressure difference across the whole channel, noted Δp , for three cycles of channel activity. Pore-pressure buildup and release varies in tandem with open/closed permeability phases, thus defining large-scale valving cycles coupled with the occurrence of seismic activity.

5.4 Random valve distributions

In order to evaluate the impact of the valve distribution on the previous results, we have generated 30 other random distributions of the same 29 valves and kept all other parameters at the same values. In all these distributions, the inter-valve distance varies randomly but there are always a number of patches where valves are closer to one another than on average. By construction, the two threshold values of the fluid flux are the same. For all these distributions, the style of activity depends on the fluid flux in exactly the same manner as for the example valve distribution represented in Figure 2b, which activity is presented in Figures 7, 8, 9, 10b, 11 and 12. For the 30 valve distributions tested, Figure 10a and c show the averaged values of the clustering exponent γ , the mean activity rate and the time-averaged proportion of open valves at each value of the input flux. We observe similar spatio-temporal patterns within individual patches and at the scale of the inter-patch distance as shown in section 5.1 for the example valve distribution (Figures 11 and 12). We conclude that characteristic patterns of synthetic activity do not depend on the specific valve distribution, but should rather depend on valve clustering in space and the average inter-valve distance in the simulated channel.

6 Discussion

The key feature of the model is that pore-pressure diffusion induced by the opening of a valve triggers neighboring valve openings. In spite of the model simplicity, the generated activity exhibits complex patterns of migration and time clustering, comparable to seismic observations. An important result is that, for a given valve distribution,

586 *i.e.* a given permeability structure of the channel, the style of activity depends on the
 587 input fluid flux.

588 In this section, we evaluate the relevance of our model in several ways. The most
 589 basic information that can be extracted from the seismic observations consists of activ-
 590 ity patterns and the associated time scales. Therefore, we first compare simulated and
 591 observed activity patterns from Guerrero, Mexico. Model results have been derived in
 592 dimensionless form and we then show that various observations are consistent with the
 593 same values for the variables in physical units. Using all these results, we draw general
 594 conclusions about the relations between the fluid flux and LFE activity in relation to the
 595 characteristics of the permeable subduction interface. We then discuss the impact of as-
 596 sumptions on valve and channel characteristics and directions for future work. In a last
 597 part, we evaluate possible extensions of the model.

598 **6.1 Comparison with LFE activity patterns at Guerrero, Mexico**

599 We first observe that activity migrates in both the updip and downdip directions
 600 within the sweet spot patch at rates that are faster than that from the sweet spot to the
 601 transient zone (Figure 1), as in our simulations. On the longest timescale, LFE activ-
 602 ity in Guerrero appears time-clustered (Frank et al., 2014, 2016). Activity bursts recur
 603 in episodic fashion and are mostly concentrated in the downdip sweet spot patch. At times,
 604 during the most vigorous episodes, activity crosses over into the less active updip tran-
 605 sient zone. Our model generates a similar pattern when the input flux q_{in} is near thresh-
 606 old values. At the lower end of the dimensionless operating flux range (*i.e.* $q_c^{break} - q_c^{clog}$),
 607 the channel is mostly closed and activity emerges when it opens up in episodic fashion.
 608 In contrast, at the upper end of the flux range, the channel is mostly open and activ-
 609 ity bursts are concentrated at the times of partial channel closure. Therefore, two dif-
 610 ferent sets of values for valve opening/closing δp thresholds, valve permeability and q_{in}
 611 could correspond to the Guerrero activity. The first situation is reminiscent of a fault-
 612 valve mechanism, as high fault zone permeability correlates in time with seismicity. Sev-
 613 eral studies interpret tremor activity in Guerrero using this correlation (Frank, Shapiro,
 614 et al., 2015; Cruz-Atienza et al., 2018). In the framework of our model, however, the sec-
 615 ond situation seems as likely as the first one. We thus argue that seismicity bursts may
 616 not be systematically linked to an increase of fault permeability and subsequent fluid pres-

617 sure surge, but may also be due to episodic phases of clogging in the fault, leading to
 618 transient fluid pressure build-up and release.

619 In our calculations, the lowermost patch is always active and is at the origin of larger
 620 scale migrations towards the less active upper patch, which trigger activity there. This
 621 is reminiscent of what happens in Guerrero (Figure 1). This behavior is a consequence
 622 of the boundary conditions that are imposed in the model. At the base of the channel,
 623 the fixed input flux constantly stimulates activity whereas the fixed pressure that is im-
 624 posed at the top acts to buffer pressure variations. Thus, activation of the upper valve
 625 cluster only occurs once sufficient fluid pressures and volumes have built up in the lower
 626 part of the channel.

627 **6.2 Scaling numerical results to LFE activity at Guerrero**

628 The main parameter to constrain is the fault-zone diffusivity $D = k/\phi\beta\eta$. Using
 629 values from the literature, we end up with a wide range which is not useful. The chan-
 630 nel porosity is $\phi = 0.01-0.05$ (Peacock et al., 2011), the fluid-rock effective compress-
 631 ibility is $\beta = 10^{-10}-10^{-9} \text{ Pa}^{-1}$ (Wibberley, 2002; Shvab & Sadus, 2015), and the dy-
 632 namic viscosity of supercritical water is $\eta = 10^{-4}-10^{-3} \text{ Pa}\cdot\text{s}^{-1}$ (Shvab & Sadus, 2015).
 633 Values for the background permeability k are very sensitive to the measurement method
 634 and, above all, to the spatial scale of the measurement. Here we do not consider laboratory-
 635 scale determinations on drill-core samples and focus on measurements that capture the
 636 large-scale channel permeability. This leads to permeability values that can be as small
 637 as 10^{-18} m^2 and as large as 10^{-11} m^2 (Doan et al., 2006; Frank, Shapiro, et al., 2015;
 638 Saffer, 2015; Hendriyana, 2021). With all these values, we find that diffusivity D lies in
 639 a very large $10^{-4}-10^5 \text{ m}^2\cdot\text{s}^{-1}$ range.

640 In order to scale model results to the Guerrero, Mexico, conditions, we need val-
 641 ues for the characteristic spatial scale \mathcal{X} , fault-zone diffusivity D , and the characteris-
 642 tic time scale \mathcal{T} . For this diffusion process, $\mathcal{T} = \mathcal{X}^2/D$. For the Guerrero tremor zone,
 643 $\mathcal{X} = 100 \text{ km}$. Using $D = 10^2 \text{ m}^2\cdot\text{s}^{-1}$, consistent with other estimates based on tremor
 644 migrations (Hendriyana, 2021), we obtain a characteristic time scale $\mathcal{T} = 10^8 \text{ s} \approx 3 \text{ years}$.
 645 Using values of the hydraulic parameters that are quoted above, a diffusivity of $D =$
 646 $10^2 \text{ m}^2\cdot\text{s}^{-1}$ implies a permeability of $k = 10^{-14}-10^{-12} \text{ m}^2$.

647 We next use the simulated activity in Figure 11 as a generic case. Such a quasi-
648 periodic behavior allows an unambiguous determination of the characteristic recurrence
649 time between large activity bursts, $T_r \approx 0.05$, and their characteristic duration, $\tau \approx$
650 0.015 , in dimensionless units. Both values are representative of highly time-clustered
651 activity, which occur for both low and high fluid fluxes, and in a large number of 29-valve
652 distributions. Scaling variables back to physical units, we find that $T_r \approx 3$ months and
653 $\tau \approx 2$ weeks, which are close to the estimates of Frank and Brodsky (2019) for Guer-
654 rero: $T_r^G = 2\text{--}3$ months, $\tau^G = 5\text{--}20$ days. In physical units, the migration velocities
655 described in section 5.2 scale up to $v = 15\text{--}20$ km.hr⁻¹ for short, rapid migrations, and
656 to $V = 7$ km.hr⁻¹ for the largest migrations that proceed from bottom to top of the
657 channel. In Guerrero, short migrations propagate indeed at velocities of a few tens of
658 kilometers per hour (Frank et al., 2014; Cruz-Atienza et al., 2018), but the larger scale
659 migrations are slower than our model estimate, with a velocity of about 1 km.day⁻¹. This
660 may be attributed to the distance that separates valve patches which has a large impact
661 on the propagation time in a diffusion process.

662 Using the above physical scales, synthetic activity rates range from 1 event per day
663 for low activity periods to about 40 events per day during bursts in near-critical fluxes,
664 and are uniformly extremely high for intermediate fluxes (Figure 10a and b). In Guer-
665 rero, average values of the activity rate are typically 600 events per day in a background
666 regime and up to a few thousands of events per day during bursts. The activity rate ob-
667 viously depends on the number of valves that are accounted for and a channel with only
668 29 valves does not accurately represent the larger-scale, more complex Guerrero system.

669 Last, but not least, we evaluate pore pressure and fluid flux values. For the fluid
670 flux scale, \mathcal{Q} , we use the estimate of van Keken et al. (2011) for the Cocos subduction.
671 The fluid flux per unit subduction zone width is $Q_T = 10$ Tg.My⁻¹.m⁻¹. Using the
672 thickness of the subduction fault zone inferred from geological observations (Angiboust
673 et al., 2015; Tarling et al., 2019), $h = 200$ m, we obtain a mass flux per unit area equal
674 to $q_T = Q_T/h = 3 \cdot 10^{-7}$ kg.s⁻¹.m⁻². This estimate is an average value for the whole
675 subduction zone and the actual fluid flow should be focused into channels at several points
676 along the strike of the fault (Ague, 2014; Piccoli et al., 2021). Such a channeling effect
677 may increase the fluid flux by up to one order of magnitude. With this caveat in mind,
678 we assume that the lower fluid flux threshold q_c^{break} , which is equal to 0.1 in dimension-
679 less units in our simulations, is of the same order of magnitude as the dehydration meta-

680 morphic flux. We deduce that $q_c^{break} \approx 3 \cdot 10^{-7} - 3 \cdot 10^{-6} \text{ kg.s}^{-1}.\text{m}^{-2}$. We then find
 681 that the characteristic pressure scale $\mathcal{P} = \eta/\rho k \mathcal{Q} \times \mathcal{X} = 30 \text{ MPa} - 30 \text{ kPa}$, using the
 682 parameter ranges previously estimated.

683 The pressure difference between the two ends of the channel Δp is close to 1 di-
 684 mensionless flux unit most of the time. Therefore, in physical units, $\Delta p \approx 30 \text{ MPa} - 30 \text{ kPa}$.
 685 Each closed valve gets unclogged for $\delta p = 0.04\mathcal{P} = 1 \text{ MPa} - 1 \text{ kPa}$. The magnitude of
 686 the force that is applied to the plug walls depends on area. For a valve in a channel with
 687 a 10^3 m^2 cross-section, Shapiro et al. (2018) were able to match the amplitude of seis-
 688 mic waves with a pressure drop of 5 MPa. Our channel might have a larger cross-section,
 689 and hence allow for large forces with comparatively lower pressure drops. The fact that
 690 these values are of the same order of magnitude demonstrates a certain self-consistency
 691 of our analysis, backed by independent observations of seismic patterns, geological flux
 692 measurements, and seismic characteristics of LFEs.

693 **6.3 Perspectives for diagnosing hydraulic conditions in real fault zones**

694 According to our model, the rate and style of seismic activity depends on the hy-
 695 draulic regime of the permeable channel. Provided that tremor and LFEs are seismic ex-
 696 pressions of valve-controlled pore pressure variations, tremor patterns can be used to in-
 697 vestigate the hydraulics of fault zones, and eventually other natural systems such as hy-
 698 drothermal fields and volcanoes where similar seismogenic mechanisms have been invoked
 699 (Honda & Yomogida, 1993)).

700 For given valve parameters that meet the condition for instability ($\delta p_c^{break}/\delta p_c^{clog} <$
 701 k_{hi}/k_{lo} , equation 16), the activity regime depends on a dimensionless input rate of fluid
 702 into the channel. This dimensionless input rate is calculated with respect to a Darcy flux
 703 for some reference permeability value (chosen to be k_{hi} here). For application to nat-
 704 ural conditions, the same dimensionless input rate can be obtained for different pairs of
 705 fluid flux and permeability values. Conversely, the same fluid flux may be responsible
 706 for different activity styles in different parts of the subduction interface depending on
 707 the local permeability value. In addition, time changes of the fluid flux, due for exam-
 708 ple to the subduction of more or less hydrated parts of an oceanic plate, may induce changes
 709 of activity.

710 The rate of fluid release by metamorphic reactions depends on the subduction rate
 711 and on the type of material that dehydrates. It cannot be the same everywhere as, for
 712 example, subduction rates vary by almost one order of magnitude. Therefore, compar-
 713 ing seismicity patterns in different subduction zones is likely to prove very useful. In par-
 714 ticular, specific attention should be paid to subduction zone segments that do not seem
 715 to generate LFE activity (*e.g.* Bocchini et al., 2021). As shown in this paper, this may
 716 be due to a fluid flux that is either very high or very small.

717 **6.4 Perspectives for exploring model configurations**

718 The present hydraulic model was designed to evaluate the potential of the clogging/unclogging
 719 valve mechanism and to illustrate the various types of activity that can be generated.
 720 It can be made more complex in many ways. As discussed above, the migration and spa-
 721 tial patterns of activity depend on the distribution of valves in the domain. The den-
 722 sity of valves can be related to the roughness of the subduction interface, which may be
 723 inherited from seafloor sedimentary cover or topography prior to subduction. Valve den-
 724 sity impacts how strongly and quickly valves can communicate through pressure tran-
 725 sients, and might be constrained by comparing magnitude-duration or magnitude-frequency
 726 scalings of synthetic and observed LFE activity. Those scalings at the LFE source level
 727 could also help constrain valve width and breaking/clogging criteria. We do not expect,
 728 however, to discover new activity regimes as solutions are determined by the four valve
 729 control parameters introduced above.

730 In all the cases reported here, all valves are identical, with the same width w , the
 731 same low permeability value k_{lo} in a state of clogging and the same values of opening/closing
 732 pressure thresholds δp_c^{break} and δp_c^{clog} . In natural conditions, the hydraulic and mechan-
 733 ical properties of the permeable fault zone should depend on the ambient pressure and
 734 hence on depth. The depth dependence of valve properties may explain — within the
 735 framework of our model — why tremor bursts recur more frequently at the deeper end
 736 of LFE areas (Wech & Creager, 2011; Frank et al., 2014). As pressure and temperature
 737 increase with increasing depth, both the background and valve permeabilities should de-
 738 crease under pressure. Depending on how permeability changes with depth, a flux that
 739 is near-critical for updip valves may well be in the intermediate range for the downdip
 740 ones. This may account for the behavior of the Guerrero tremor area, where activity is
 741 much more frequent and continuous downdip than updip. With lower permeability, mi-

742 grations should also be slower. With that in mind, the fast migration rates observed in
 743 the sweet spot in Guerrero would imply that valve density might increase in this depth
 744 range.

745 Model parameters that must be explored include the valve characteristics (length,
 746 permeability in a state of clogging), the distribution of valves and the total number of
 747 valves. However, the valve instability criterion (equation 16) constrains how permeabil-
 748 ity and pressure threshold values can vary. As discussed above, we expect that this con-
 749 dition is met in the pore-pore throat configuration that is relevant to many natural per-
 750 meable systems (McDowell-Boyer et al., 1986; Beckingham, 2017). Ultimately, a more
 751 realistic description of the process should include time-dependent opening and closure
 752 processes, but very few constraints on the associated time-scales are available.

753 We have shown in section 5.3 that neighboring valves may act as a single macro-
 754 valve in some conditions (Figure 12), and this is worthy of a dedicated investigation. De-
 755 termining which valve spacing and hydraulic conditions enable this behavior is likely to
 756 provide key insight into the interaction mechanisms that generate large bursts of activ-
 757 ity. One could also allow for some random distribution of valve properties. However, we
 758 may draw from our current understanding and predict that the end result would be that
 759 only a subset of valves get activated for a given flux. Other boundary conditions than
 760 those used here are also worth investigating. For example, the subduction channel may
 761 not be connected to a very large permeability medium and allowing for an output flux
 762 that depends on pressure may lead to large-scale cyclic activity. In addition, dehydra-
 763 tion surges may occur due for example to kinetic constraints on metamorphic reactions
 764 or to highly hydrated portions of the downgoing plate crossing the phase-change bound-
 765 ary. Ongoing and future work with similar models will investigate these mechanisms and
 766 further explore parametric controls on valve behavior.

767 **6.5 Coupling between hydraulic transients and fault slip**

768 An important direction to explore in the future is the coupling between hydraulic
 769 transients and fault slip. The latter can be described using a RSF friction parameter-
 770 ization with coefficients that depend on pore pressure (*e.g.* Luo & Liu, 2019). In such
 771 a framework, the main challenge is to constrain the causal relationship between fluid pres-
 772 sure transients, in-channel damage (valve opening), and slip. A key factor is a high pore

773 pressure because it lowers the fault strength. As discussed in the introduction, most mod-
 774 els of RSF fault slip consider that slip comes first and induces a permeability increase
 775 through fracturing and dilation. Another perspective to consider is the effect of changes
 776 of permeability on frictional stability, which would allow to explore how slip could be
 777 influenced by valving in the fault. Unclogging acts to lower pore pressures and hence to
 778 increase the fault strength. At the same time, it lowers the effective friction coefficient
 779 because it reduces the solid to solid contact area, *i.e.* the roughness of the fault inter-
 780 face. In this manner, unclogging may induce slip. The fluid pressure component of fault
 781 strength may therefore be modulated by the openness of valves because closed valves act
 782 as contact points between fracture walls and hence may act as high effective normal stress
 783 barriers to slip. The interactions between fault slip, damage and permeability evolution
 784 should be explored in depth in coupled physical models.

785 **7 Conclusion**

786 In this work, we have developed a model to explain the tremor patterns in subduc-
 787 tion zones as symptoms of rapid, strong, localized pressure transients emerging from in-
 788 termittent fluid transport in a permeable fault interface. We model fluid transport in
 789 a 1D permeable channel, fed by a constant input flux. Based on field observations in faults,
 790 and laboratory experiments in porous media, we have argued that clogging and un-clogging
 791 events occur in the permeable channel, causing large changes of permeability. Based on
 792 simple physical and geological arguments, we have shown that such events can alternate
 793 in a simple valve mechanism. The strong pressure transient that occurs when a valve opens
 794 is able to generate a seismic event, and we have therefore considered valve activity as
 795 synthetic seismic activity.

796 We have found that the behavior of a permeable channel containing valves depends
 797 on two valve parameters (pressure ratio $\delta p_c^{break}/\delta p_c^{clog}$ and permeability ratio k_{hi}/k_{lo}).
 798 In one case ($\delta p_c^{break}/\delta p_c^{clog} > k_{hi}/k_{lo}$), the channel is in a stable steady-state regime
 799 with no valve activity for all values of the input flux. In the other case ($\delta p_c^{break}/\delta p_c^{clog} <$
 800 k_{hi}/k_{lo}), the channel is unstable with valves that open and close repeatedly if the input
 801 flux is within a well-defined operating range. We have shown that only the latter is rel-
 802 evant to geological systems. Therefore, in this framework, the value of the fluid flux go-
 803 ing through a fault zone is a major control on the observed seismicity. For values of the
 804 flux outside the operating range, the fault zone is permanently closed or open. For val-

805 ues within the operating range, permeability valves opening and closing in cycles gen-
 806 erate permanent activity and intermittent fluid transport. In this permanently unsta-
 807 ble regime, spatiotemporal patterns of modeled tremor emerge from valve-to-valve in-
 808 teractions, and are very similar to those observed in many subduction zones.

809 Within the operating range for permanent activity, the input flux controls the dy-
 810 namics of the permeable system. The system is mostly closed for low flux values and mostly
 811 open for high values, with time-clustered bursts of activity. For intermediate flux val-
 812 ues within the range, the system is mostly open, and closes in quasi-periodic bursts of
 813 activity. Furthermore, the activity rate is highest and almost constant through time. Us-
 814 ing this model and more elaborate versions of it, one may hope to use microseismicity
 815 patterns to probe the hydraulics of permeable fault zones and how they change with time.

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 822 is made available on the Slow-Earthquake database, under the name Frank2014 ([http://www-
 823 solid.eps.s.u-tokyo.ac.jp/sloweq/](http://www-solid.eps.s.u-tokyo.ac.jp/sloweq/)).

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Table 1. Characteristic dimensions used to scale physical variables

Variable		Characteristic scale	
x	Distance along channel	$\mathcal{X} = L$	Channel length
z	Depth	$\mathcal{Z} = \mathcal{X} \sin(\alpha)$	Channel vertical depth
t	Time	$\mathcal{T} = \mathcal{X}^2/D$	Diffusive time ^a
p	Pore pressure	$\mathcal{P} = (\rho_r - \rho)g\mathcal{Z}$	Dynamic pressure difference across channel
q	Massic flow rate ^b	$\mathcal{Q} = k_{hi}\rho/\eta \times \mathcal{P}/\mathcal{X}$	Flux resulting from characteristic pore pressure difference ^a
k	Permeability	$\mathcal{K} = k_{hi}$	Background channel permeability

^a along a fully open channel, $D = k_{hi}/(\rho\eta\beta)$

^b per unit channel cross section area

Table 2. Valve parameters for all simulations presented in this study

Parameter		Value (scaled)
w	Valve width	0.02
k_{hi}	Open valve permeability	1
k_{lo}	Closed valve permeability	0.05
δp_c^{break}	Threshold p difference for opening	0.04
δp_c^{clog}	Threshold p difference for closing	0.021
q_c^{break}	Flux above which a closed valve can open	0.1
q_c^{clog}	Flux below which an open valve can close	1.05

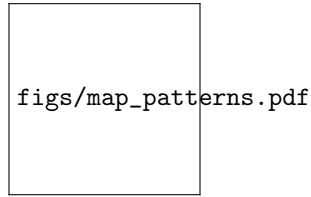


Figure 1. The Guerrero (Mexico) tremor zone in (a) map view and (b) vertical cross-section. Each dot represents a low-frequency earthquake family from Frank et al. (2014). Thick black lines outline the trench and the subduction interface (data from Hayes (2018)). (c) and (d), LFE activity in a time-location chart from the Frank et al. (2014) catalog. Activity occurs in two patches centered on the subduction interface, a dense downdip one called the *sweet spot* and a sparser updip one called the *transient zone*. Tremor activity migrates along the subduction interface in both the updip and downdip directions, as shown by the orange arrows.

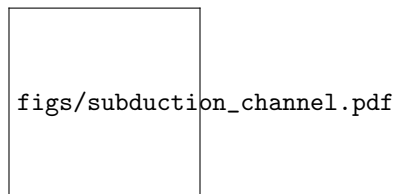


Figure 2. (a) Schematic representation of a permeable fault zone. Below the tremor source region, metamorphic dehydration reactions release fluid, which is channeled in the permeable subduction interface as it circulates under high fluid pressures. (b) Permeability structure for a 1-D hydraulic model of this system, involving a number of randomly distributed plugs that can open and close under certain conditions. The channel is fed at its base by input fluid flux q_{in} . At the top, the channel is connected to a high permeability region such that the fluid pressure is close to the hydrostatic pressure p_{out} . (c) Initial conditions correspond to steady-state flow regime with a small input flux and all plugs closed.

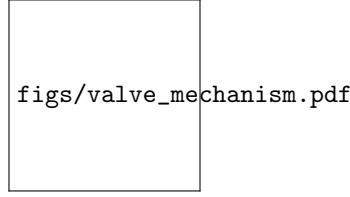


Figure 3. (a) The 4 stages of the valve mechanism. 1, the valve has been closed and permeability is k_{lo} . δp , the pressure difference across the valve, increases. 2, when δp reaches the threshold value δp_c^{break} , an impulsive unclogging event (an erosive burst) occurs. Particles get flushed out, the valve opens and permeability rises to k_{hi} . 3, valve opening generates a rapid pressure transient that is associated with a low frequency earthquake. 4, when δp decreases to δp_c^{clog} , the deposition of fines and colloids resumes and permeability drops to k_{lo} . The valve closes and a new cycle begins. (b) Hysteretical cycle of pressure difference δp and valve permeability k . (c). Evolution of δp through 3 successive cycles of valve opening and closure for an isolated valve.

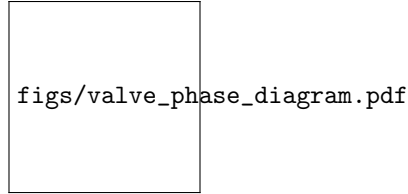


Figure 4. Requirements for steady-state valve regimes (equations (14) and (15)) in k - δp space (in log-log scale). The two values for permeability k and threshold values for the valve pressure difference δp are shown with vertical and horizontal dashed lines. In this space, iso-flux contours appear in gray lines of slope -1 in log-log scale (-45°). Blue and red segments indicate values of δp that are allowed for a closed valve and an open one in steady-state, respectively. Steady-state conditions are achieved in the green and red shaded areas. See explanations in the text. (a) Valve with enhanced permeability variations ($k_{hi}/k_{lo} > \delta p_c^{break}/\delta p_c^{clog}$). (b) Valve with enhanced pressure variations ($k_{hi}/k_{lo} < \delta p_c^{break}/\delta p_c^{clog}$).

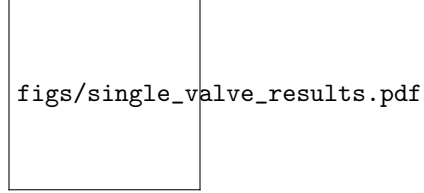


Figure 5. Influence of the input flux q_{in} on the activity of a single valve. (a) Three intervals of input flux q_{in} are delimited by two threshold values q_c^{clog} and q_c^{break} . (b) Examples of activity rate evolution in time (in black) for each regime. Activity rate is taken as the number of events per $\delta t = 0.01$, scaled unit. An event corresponds to the opening of a valve. Valve parameters are listed in Table 2.

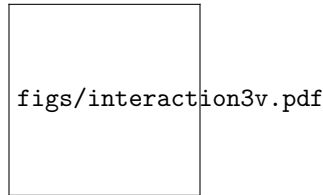


Figure 6. Interactions between three neighboring valves. (a) Transient evolution of pore pressure from an initial equilibrium profile with all valves closed towards a profile with valves $v1$ and $v3$ closed and valve $v2$ open. Panels (b) and (c): pore pressure difference evolution across valves $v1$ and $v3$.

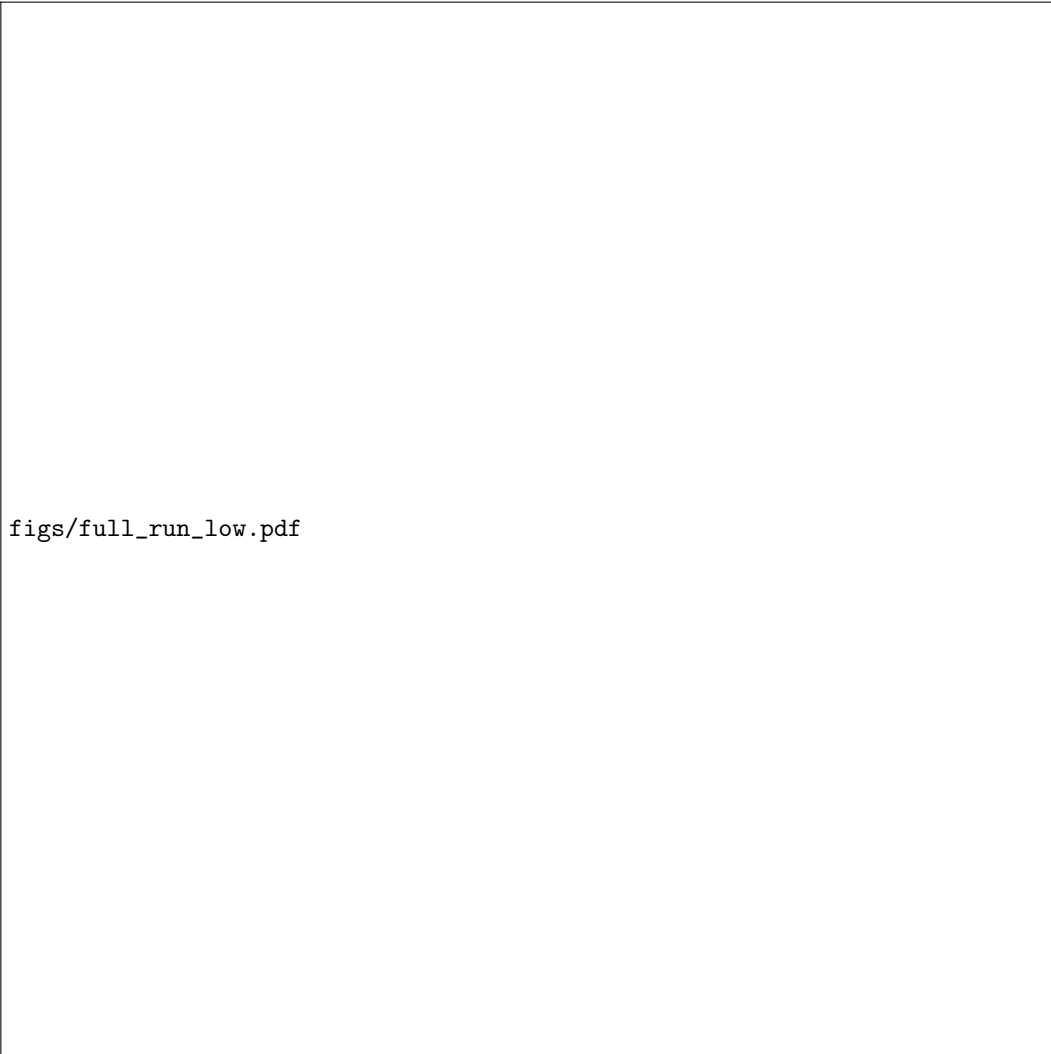


Figure 7. Time evolution of activity for a channel with 29 randomly distributed valves and a low input flux $q_{in} = 0.16$, just above q_c^{break} . (a) Fluid pressure differential across the channel Δp , (b) equivalent permeability k_{eq} and (c) activity rate (events per $\delta t = 4 \times 10^{-4}$ scaled unit). Raw data in grey and smoothed time series in colors (rolling mean with a window of length $\tau = 0.03$). (d) Time-space diagram of events occurrence within the channel. All valves are initially closed. A permanent regime with intermittent activity is reached after a short initial transient. Valve distribution represented in Figure 2, valve parameters are listed in Table 2.

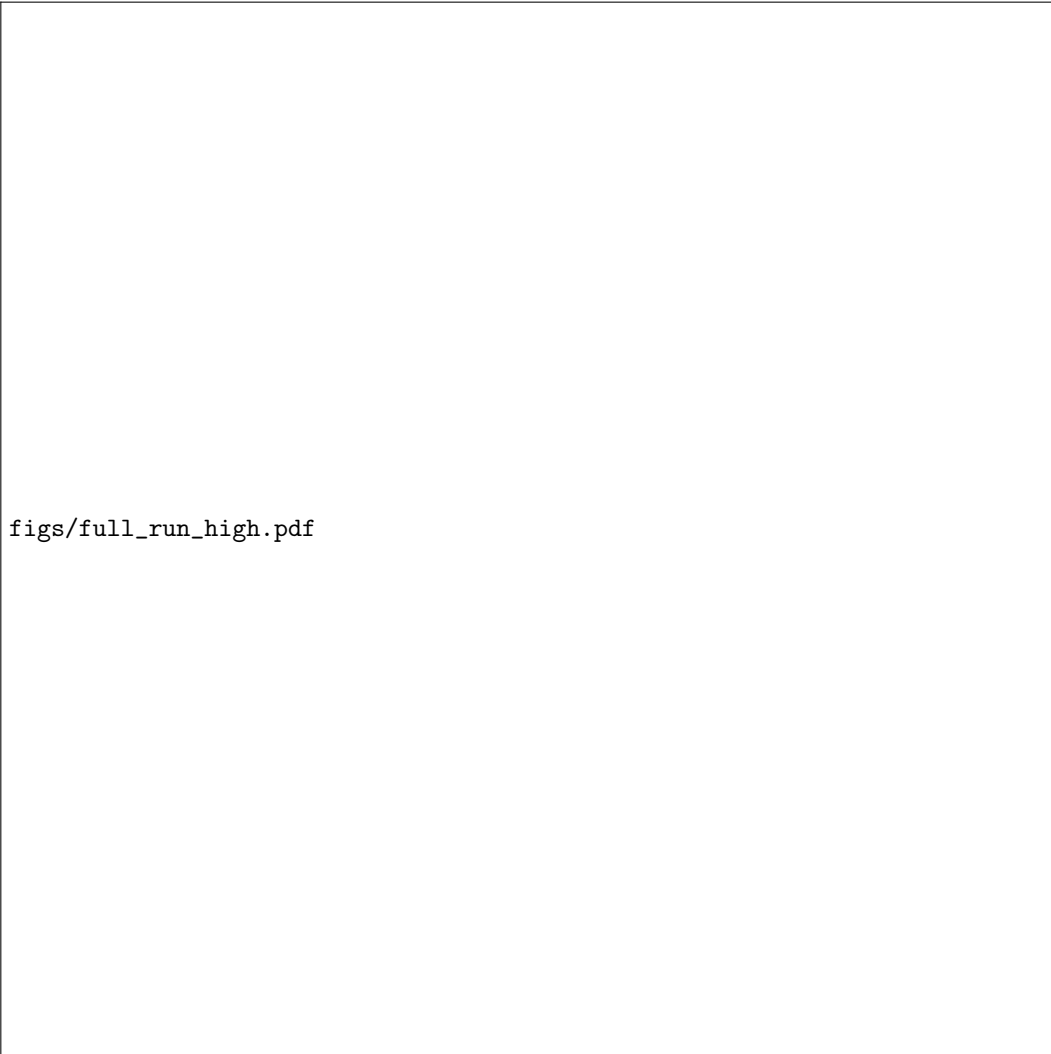


Figure 8. Time evolution of activity for a channel with 29 randomly distributed valves and a high input flux $q_{in} = 0.81$, just below q_c^{clog} . (a) Fluid pressure differential across the channel Δp , (b) equivalent permeability k_{eq} and (c) activity rate (events per $\delta t = 4 \times 10^{-4}$ scaled unit). Raw data in grey and smoothed time series in colors (rolling mean with a window of length $\tau = 0.03$). (d) Time-space diagram of events occurrence within the channel. All valves are initially closed. A permanent regime with intermittent activity is reached after a short initial transient. Valve distribution represented in Figure 2, valve parameters are listed in Table 2.

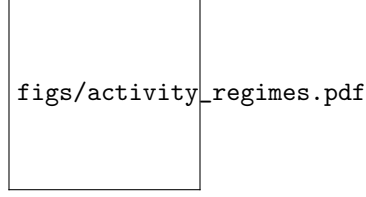


Figure 9. Influence of the input flux q_{in} in a representative system with 29, randomly distributed valves. (a) Three intervals of input flux q_{in} are delimited by two threshold values q_c^{clog} and q_c^{break} . (b) Examples of activity rate (number of events per $\delta t = 5 \times 10^{-4}$, scaled unit, in black) and proportion of open valves (in cyan) as a function of time, starting at $t = 0$ with all valves closed. Valve distribution represented in Figure 2, valve parameters are listed in Table 2.



Figure 10. Different permanent activity styles as a function of the input fluid flux, averaged for 30 different 29-valve systems. (a) In white, the level of time clustering γ : the spectral slope of autocorrelation of event counts (see text) ; in gray, mean activity rate: number of events per unit time, both as a function of input flux q_{in} . For γ close to 0, activity is Poisson-like with no time clustering. The higher γ is, the more clustered and episodic activity is. (b) Activity rate (in black) and proportion of open valves (in cyan) as a function of time for particular simulations within the different regimes. (c) Time-averaged proportion of open valves in simulations, as a function of input flux. In (a) and (c), each dot represent the average value of the variable over 30 different 29-valve systems, at the given flux. Error bars are present in both panels, representing the standard deviation of this measurement over the 30 systems.

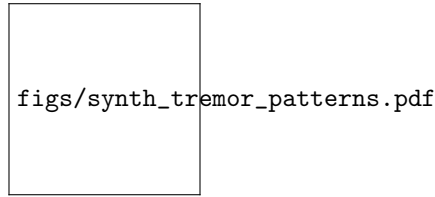


Figure 11. Synthetic tremor activity patterns. *(a)* Distribution of valves in the permeable channel. This distribution was generated randomly and contains three large clusters. *(b)* Activity patterns in space and time. Black dots stand for valve opening events. The lower valve cluster ($x = 0 - 0.18$) is active almost continuously. The middle and upper clusters ($x = 0.2 - 0.65$ and $x = 0.7 - 1$) is activated sequentially in bursts. Within a burst, activity is intermittent too. *(c)* focus on the activity in the middle valve cluster showing fast, short-scale migrations in both the updip and downdip directions. Valve distribution represented in Figure 2, valve parameters are listed in Table 2.

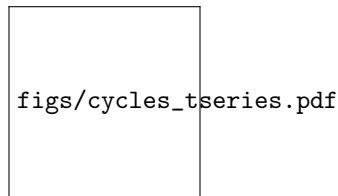


Figure 12. Cyclic variations of activity for $q_{in} = 0.93$. *(a)* Pore pressure difference across the whole channel Δp , *(b)* channel bulk permeability k_{eq} , and *(c)* time-location chart of events. *(d)* Activity in $(\Delta p, k_{eq})$ space, showing hysteretical cycles like those of a single valve. Valve distribution represented in Figure 2, valve parameters are listed in Table 2.

Figure 1.

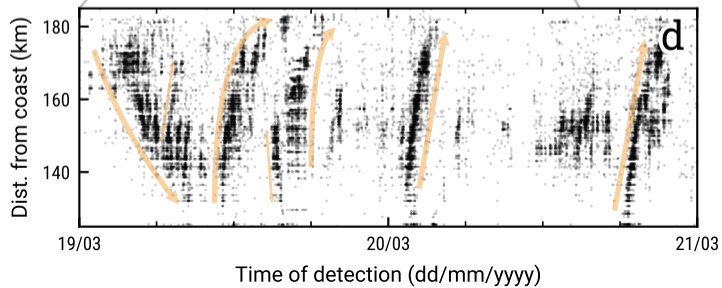
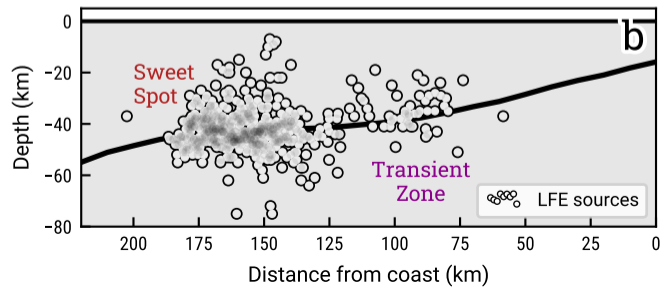
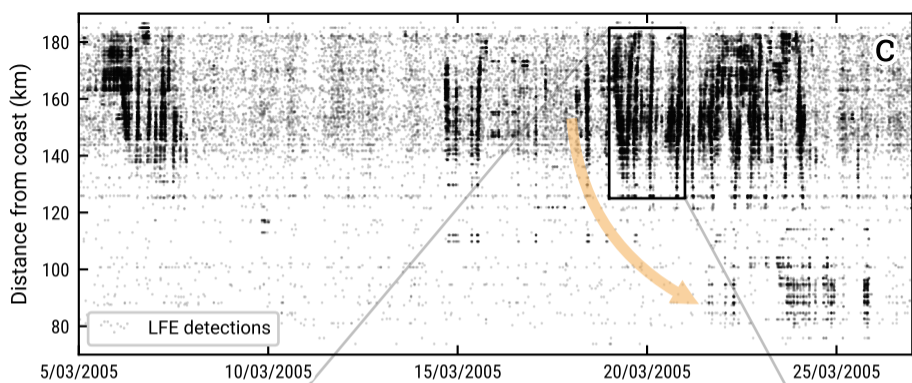
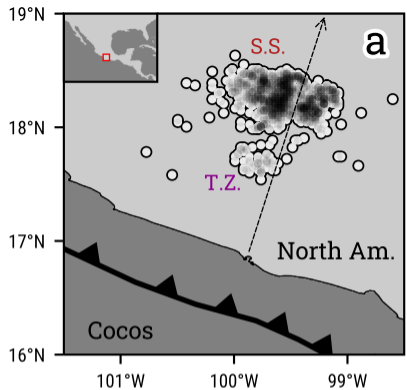


Figure 2.

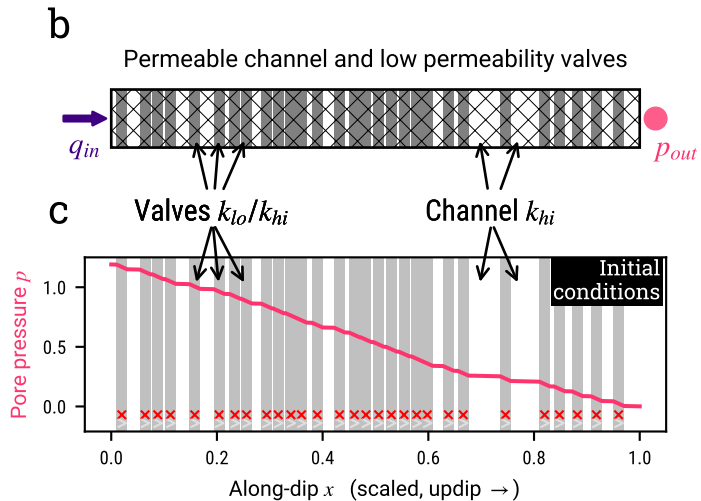
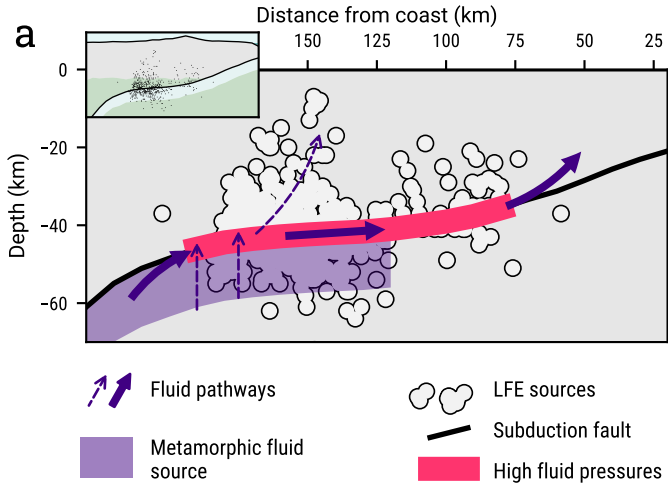


Figure 3.

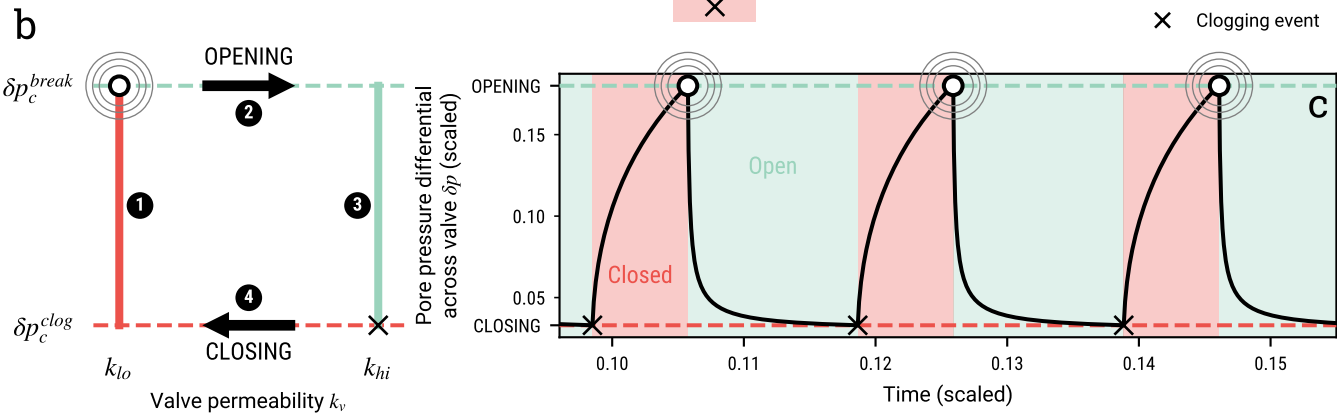
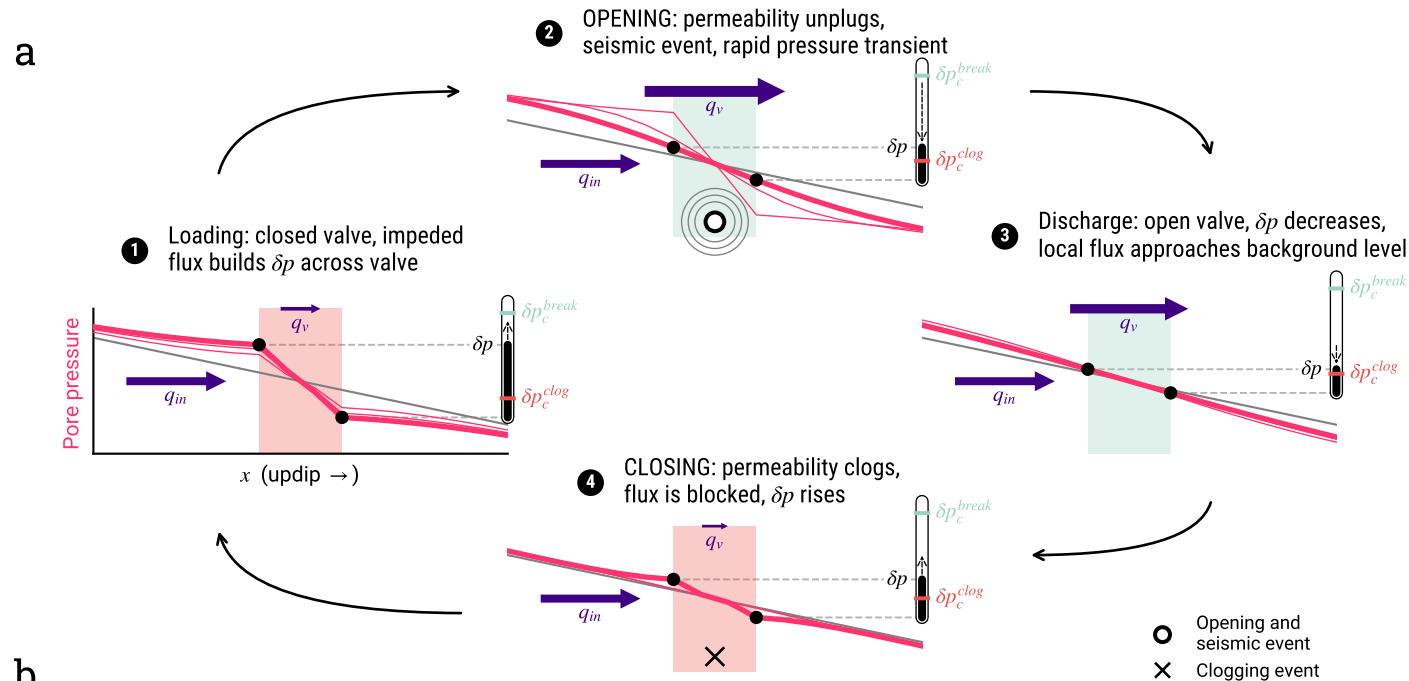


Figure 4.

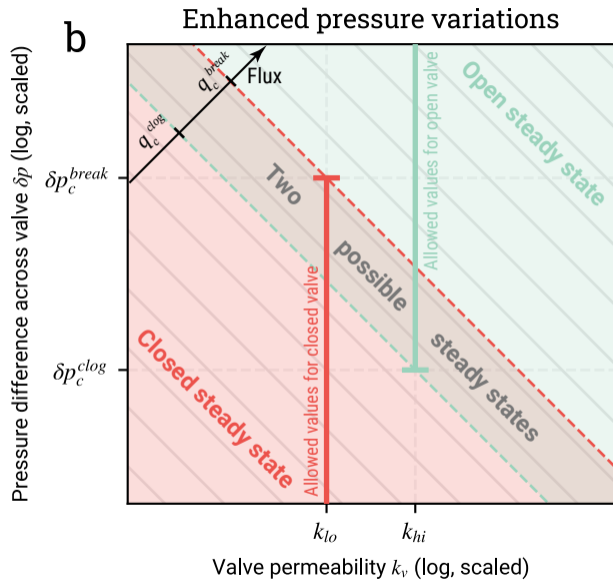
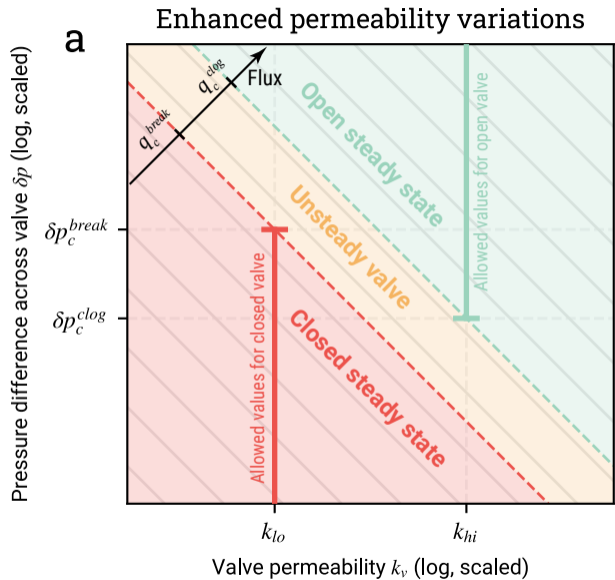
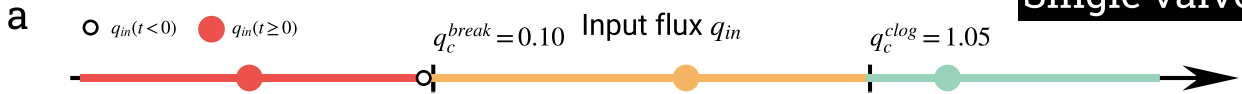


Figure 5.

Single valve



b

Equilibrium: valve stays closed **Permanent activity: valve opens and closes** **Equilibrium: valve stays open**

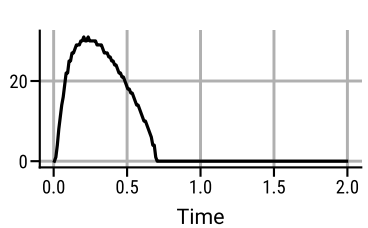
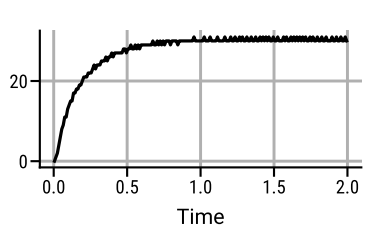
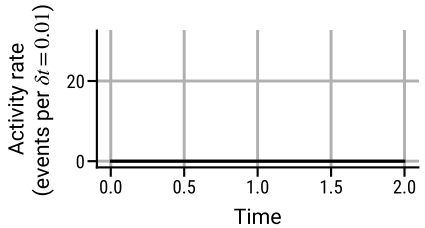


Figure 6.

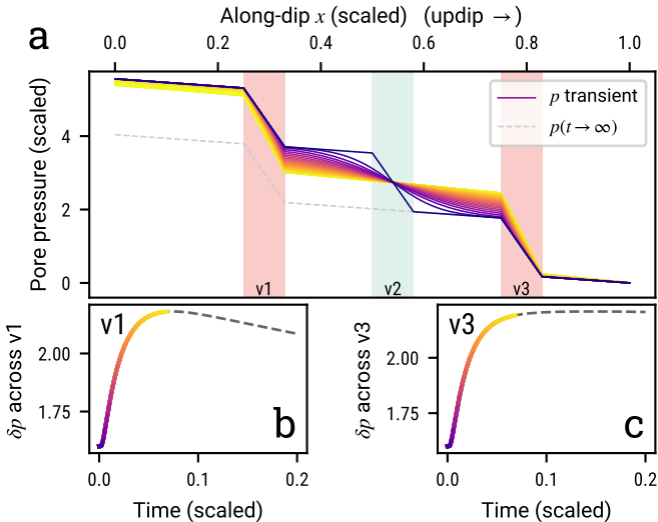


Figure 7.

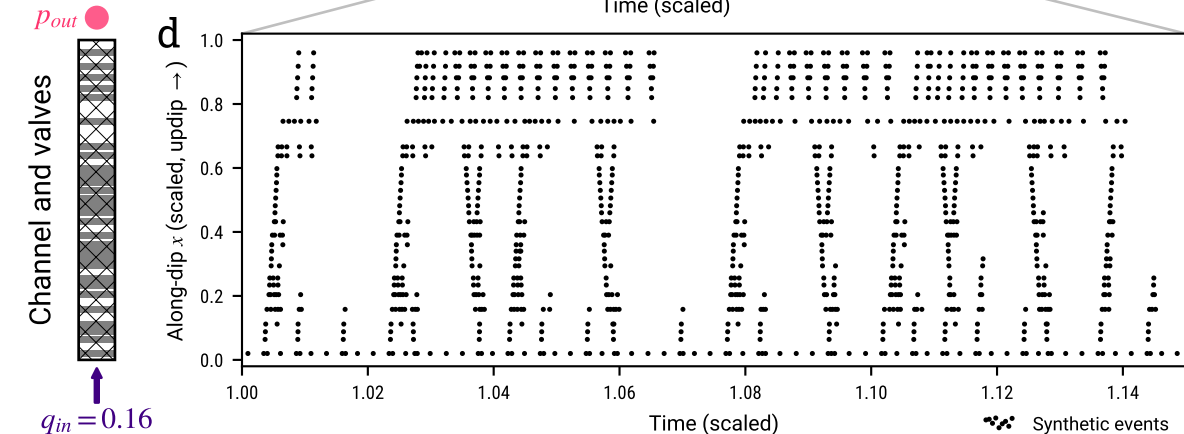
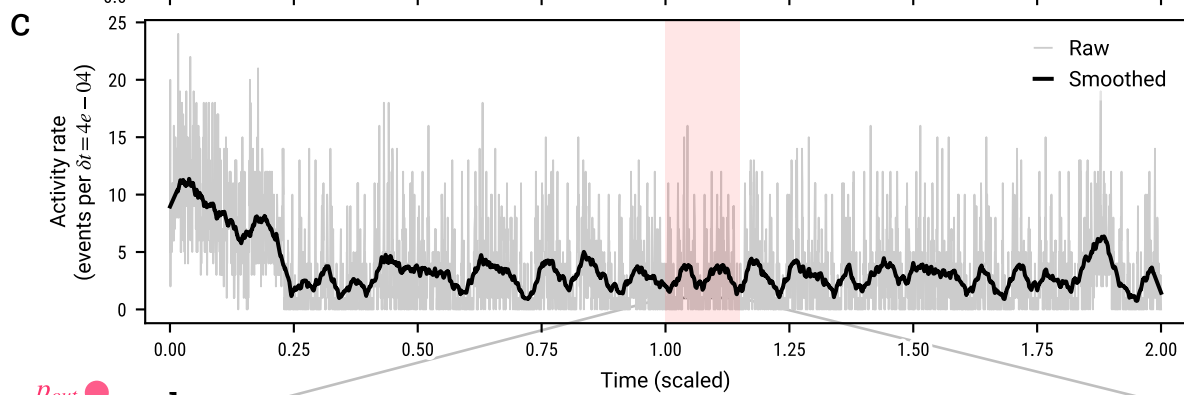
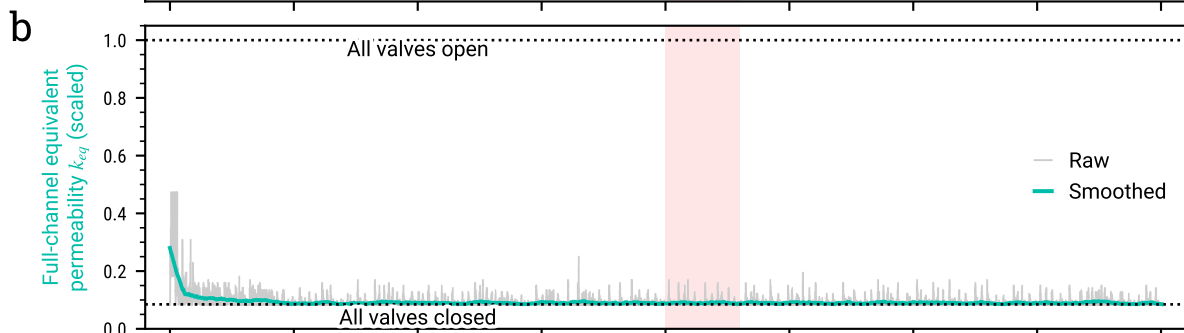
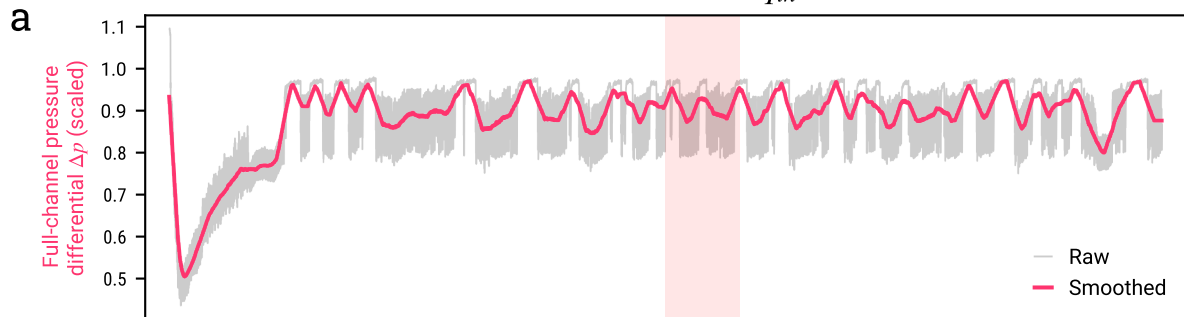
Simulation results: $q_{in} = 0.16$ 

Figure 9.

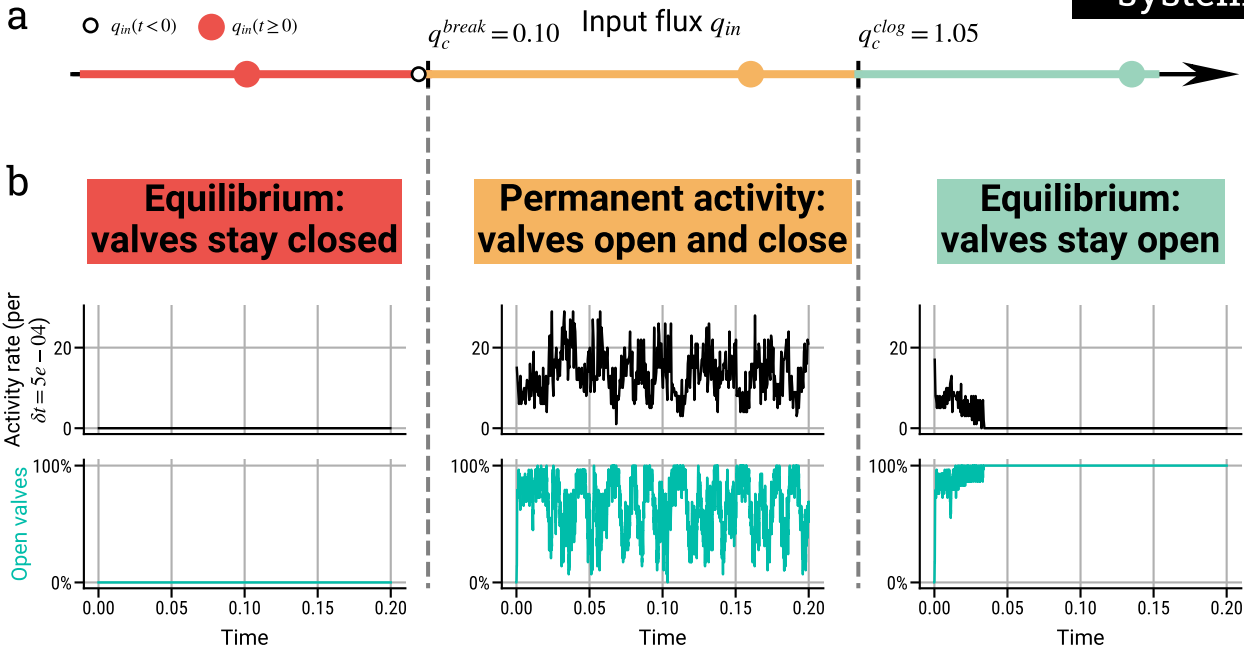


Figure 8.

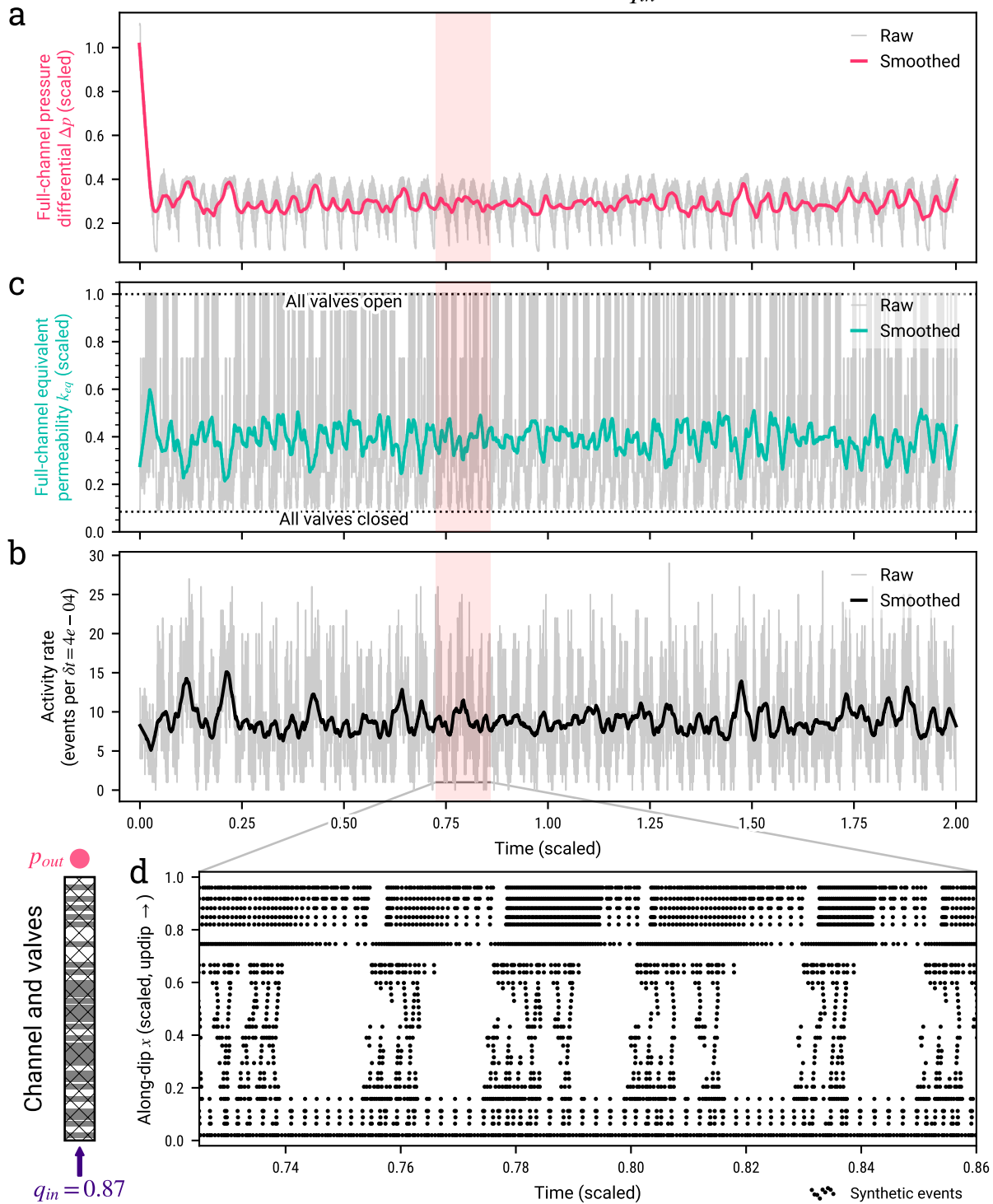
Simulation results: $q_{in} = 0.87$ 

Figure 10.

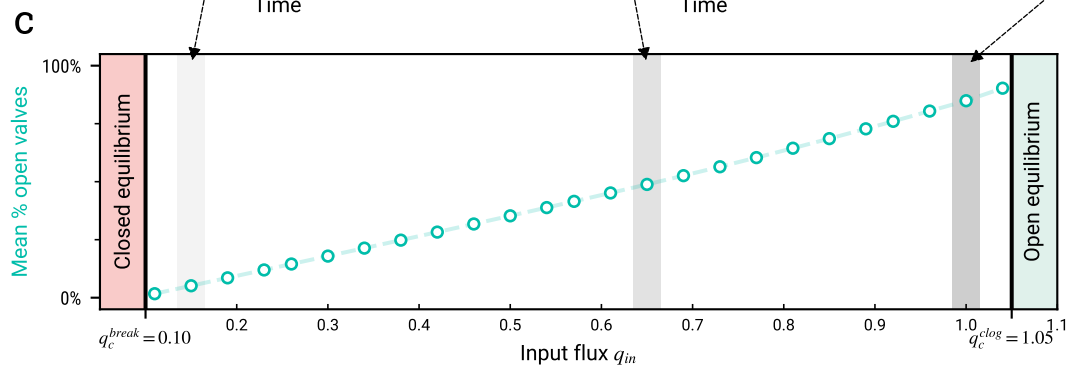
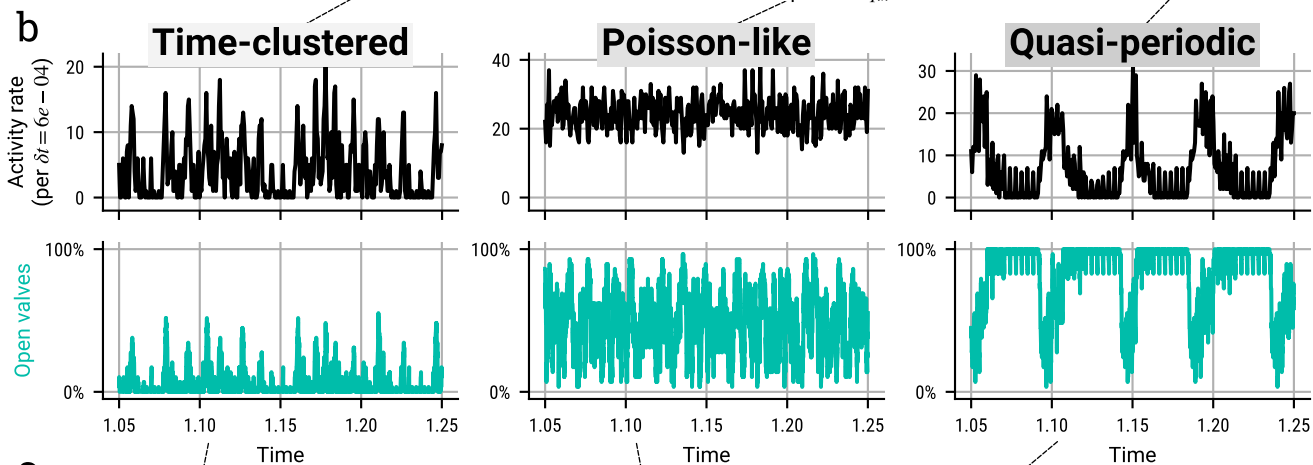
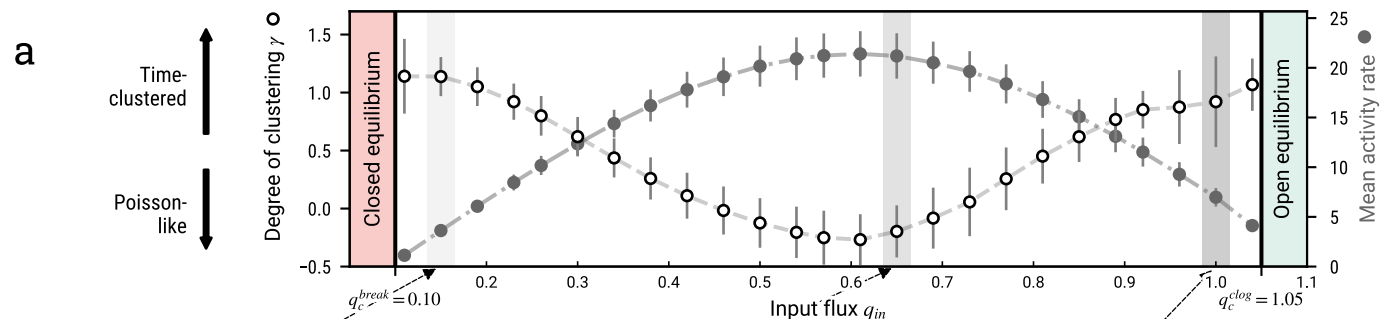


Figure 11.

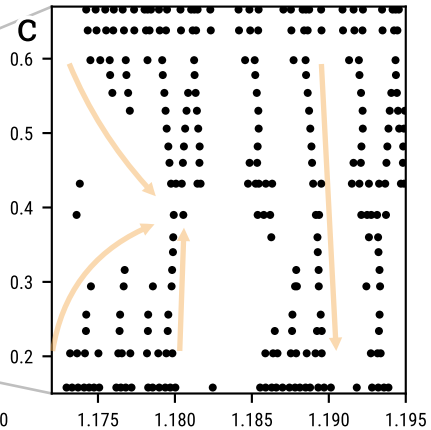
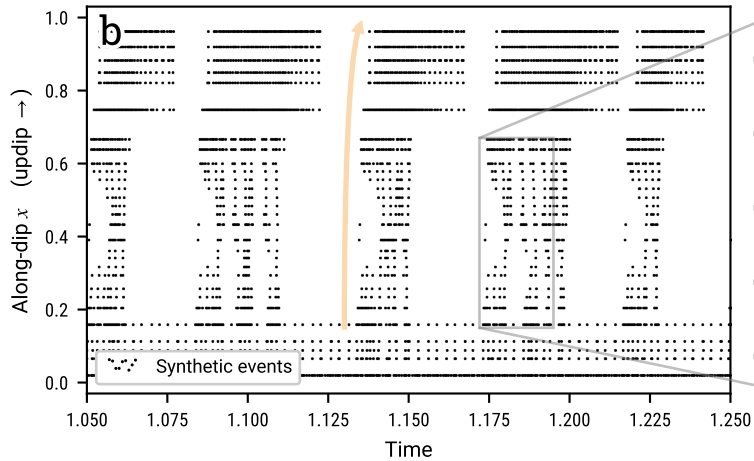
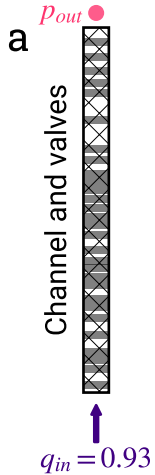
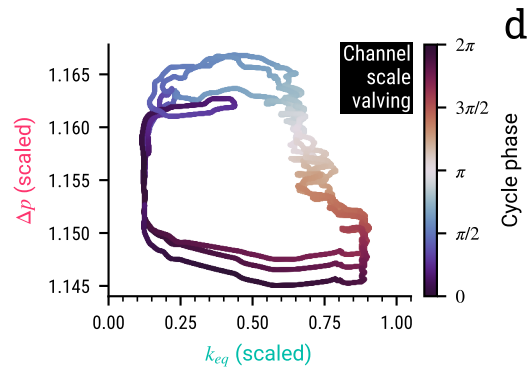
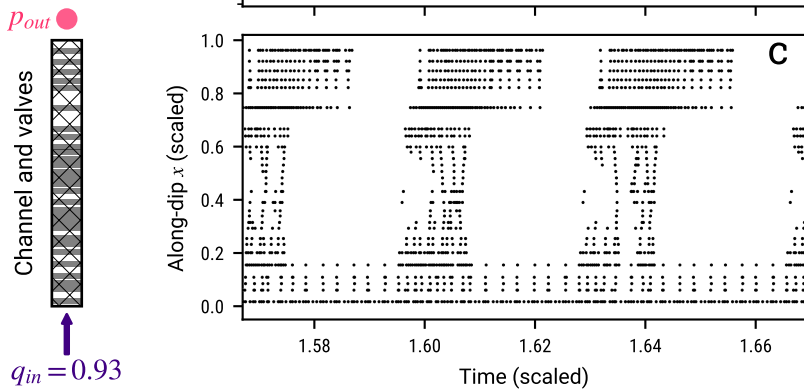
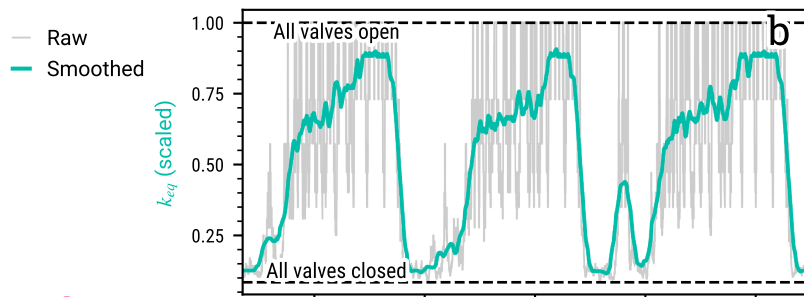
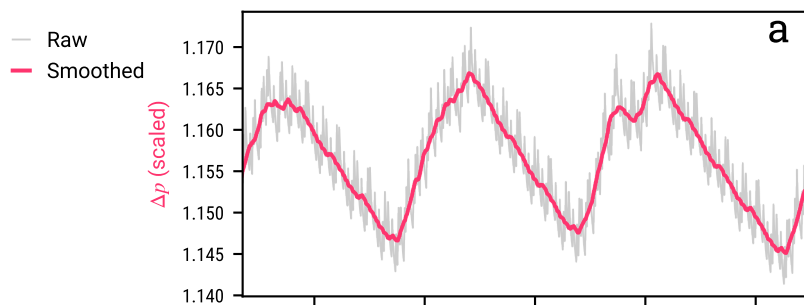


Figure 12.



⋯ Synthetic events