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### 2 Self-induced seismicity due to fluid circulation along faults.

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11

#### 12 Abstract

13 In this article, we develop a system of equations describing fluid migration, fault 14 rheology, fault thickness evolution and shear rupture during a seismic cycle, triggered either 15 by tectonic loading or by fluid injection. Assuming that the phenomena predominantly take 16 place on a single fault described as a finite permeable zone of variable width, we are able to 17 project the equations within the volumetric fault core onto the 2D fault interface. From the 18 basis of this "fault lubrication approximation", we simulate the evolution of seismicity when 19 fluid is injected at one point along the fault to model induced seismicity during an injection 20 test in a borehole that intercepts the fault. We perform several parametric studies to 21 understand the basic behaviour of the system. Fluid transmissivity and fault rheology are key 22 elements. The simulated seismicity generally tends to rapidly evolve after triggering, 23 independently of the injection history and end when the stationary path of fluid flow is 24 established at the outer boundary of the model. This self-induced seismicity takes place in the 25 case where shear rupturing on a planar fault becomes dominant over the fluid migration 26 process. On the contrary, if healing processes take place, so that the fluid mass is trapped 27 along the fault, rupturing occurs continuously during the injection period. Seismicity and fluid 28 migration are strongly influenced by the injection rate and the heterogeneity.

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### 1 1. Introduction

2 Some seismicity is believed to be driven by fluid circulation, because high-pressure fluid in 3 the fault zone can reduce frictional strength. In principal, the following two categories can be 4 identified. The first of these is naturally-occurring induced seismicity, including volcanic 5 seismicity, some aftershocks of large earthquakes, or seismicity in subduction contexts where large quantities of fluid can be expected to be present. In such examples, although the 6 7 existence of the fluid can be imaged from seismic tomography, it is very difficult to quantify 8 fluid migration and the resultant seismicity. The second category is identified as 9 anthropogenically-induced seismicity, forced by industrial or other forms of manmade 10 injection or extraction of fluid such as the extraction of natural gas, CO<sub>2</sub> storage and the 11 development of deep geothermal systems.

12

13 Many cases of naturally-occurring induced seismicity have been reported over recent 14 decades. One notable example was the Matsushiro earthquake swarm that lasted two years 15 starting in 1965 and featured more than 60 000 felt earthquakes, the largest of which had a 16 magnitude of 5.4. Although various models based on dilatancy and magma-intrusion had been 17 proposed, the events can more probably be ascribed to a massive migration of fluids including 18 outflow to the ground (e.g., Ohtake, 1976; Matsu'ura and Karakama, 2005; Cappa et al., 19 2009). Some seismic activities triggered in the wake of large earthquakes suggest the 20 important role of fluids, as demonstrated as far away as 1250 km from the epicentral zone in 21 the wake of the 1992 Landers, California, earthquake (Hill et al., 1993). Such triggered 22 earthquake swarms have been observed for other earthquakes too, such as the Yalova cluster 23 that ensued after the 1999 Izmit earthquake (Karabulut et al., 2011) or regional seismicity 24 following the 1997-1998 Umbria-Marche, Italy, sequence (Lombardi et al., 2010). 25 Earthquake migrations have also been observed as a result of the release of CO<sub>2</sub>-rich water 26 (Miller et al., 2004). Some crustal earthquakes may have been triggered due to high pore 27 pressure, as inferred for the 2009 L'Aquila, Italy, earthquake (Terakawa et al., 2010). 28 Furthermore, the importance of the existence of fluids has been emphasized for plate 29 boundaries, in particular for subduction associated with microseismicity or aseismic slip. 30 Obara (2002) reported that the discovered deep, non-volcanic tremors along the subducting 31 Philippine Sea plate may be related to the fluid generated by dehydration processes from the 32 slab. In many cases, the existence of fluid is inferred from seismic tomography as an anomaly 33 of V<sub>p</sub>/V<sub>s</sub>, the ratio of P-wave and S-wave velocities (O'Connell and Budiansky, 1974;

Thurber *et al.*, 1997 and others). Seismicity clusters attributed to fluid migration have also
 been observed in stable tectonic contexts, such as Remiremont, France (Audin et *al.*, 2002).

3

4 On the other hand, it is recognized that seismicity can be induced directly in 5 conjunction with fluid injection (Shapiro et al., 1999, Shapiro and Dinske, 2009), in particular, 6 in deep geothermal projects (Person, 1981; Cornet and Jianmin, 1995; Calò et al., 2011) or 7 CO<sub>2</sub> injection projects (Cappa and Rutqvist, 2011a, 2012; Zoback et al., 2012; Mazzoldi et al., 8 2012). The stimulation phases of the deep geothermal projects, in particular, require 9 seismogenic reactivation of pre-existing fractures or micro-fracturing to allow fluid circulation in the targeted reservoir. Assessing this type of seismicity is, therefore, becoming 10 11 an important issue. For example, at Soultz-sous-Forêts (Alsace, France), an Enhanced 12 Geothermal System (EGS), has been monitored since an initial borehole was drilled to a depth 13 of 2000 m in 1987 (Kappeimeyer et al., 1991; Beauce et al., 1991; Fabriol et al., 1994; Evans 14 et al., 2005; Neuville et al., 2009, Gentier et al., 2011a, b; Evans et al., 2012). Subsequently, 15 deeper wells were drilled to depths of about 5000 m, and several stimulations were conducted 16 at regular intervals after 2000 (Dorbath et al., 2009). An injection experiment typically lasts a 17 few days (about 100-250 hours), with a maximum injected fluid volume speed of 50 L/s and wellhead pressure of 17 MPa. The total volume of injected fluid amounts to about 20 000-30 18 19 000 m<sup>3</sup>. More than 100 000 seismic events were detected in the course of the three stimulation 20 experiments, with rates up to 8000 events per day (Baisch et al., 2010). Typically, the 21 seismicity is localized as a cloud surrounding the injection point, which spatially expands 22 with time (Shapiro et al., 1999). Sometimes a quiet zone appears around the injection point 23 once it has been sufficiently stimulated, while the fluid and seismicity migrate outside. This is 24 known as the Kaiser (1950) effect. Although the seismicity briefly expands in a complex 3D 25 medium, the relocated earthquake locations seem to align along one or more planes, construed 26 to be pre-existing faults at the site (Dorbath et al., 2009; Baisch et al., 2010).

27

Naturally-triggered or anthropogenically-induced seismicity models have been developed by various researchers in seismology, rock mechanics and other areas of specialization. For injection experiments, volumetric models (sometimes containing networks of linear or plane fractures) have been considered (Bruel, 2002, 2007; Gentier *et al.*, 2011a, 2011b). However, if the seismicity occurs predominantly along some pre-existing fault(s), a linear fault (in 2D) or a plane fault (in 3D) are often studied (e.g., Blanpied *et al.*, 1992; Segall and Rice, 1995; Baisch *et al.*, 2010; Cappa and Rutquist, 2011b, 2012). Fault rheology

1 is a key element allowing fluid transport. Early models of fluid-driven seismicity triggering 2 simply considered the pressure balance in broken portions to be instantaneous, corresponding 3 to an assumption of very high porosity in these sections (Miller et al. 1996). Since natural materials mostly give rise to low Reynolds numbers, due to their low porosity, more precise 4 5 models incorporate a finite viscosity and pore-fluid transport through permeable parts 6 (governed by Darcy's law, e.g., Walder and Nur, 1984). When finite compressibility of the 7 fluid plays a role, it is reported that this leads to associated pore pressure diffusion (Shapiro et al., 1999, Goren et al., 2010, 2011). This compressibility is shown to be involved in most 8 9 systems of shear faults, and must be taken into account to understand the evolution of pore 10 pressure around the fault (Goren et al. 2010).

11

12 Experimentally, fracturing and finger propagation during fluid injection in analogue 13 faults have been observed at the laboratory scale, in transparent impermeable cells filled with 14 granular materials comparable to fault gouge (Johnsen et al., 2006, 2007, Cheng et al. 2008, 15 Huang et al. 2012a, 2012b). Similar pattern formations were observed after injection of a 16 slightly compressible and viscous fluid (oil), and of a compressible and slightly viscous 17 substance (air) (Johnsen et al., 2008) – as was the case for the formation of decompaction 18 fronts in such systems (Vinningland et al., 2012). The fracturing and fingering triggered by 19 fluid injection was shown experimentally to be accompanied by microseismicity (Schelstraete, 20 2009).

21

22 To address this type of fluid-driven fracturing, modelling the momentum exchange 23 between the flowing fluid and the elastic solid is essential. When the two are considered as 24 distinct, continuous bodies, mixture theories have been developed that are widely used to 25 model wet landslides (Iverson, 1997). In the context of high fluid flow in highly deformable 26 solids, similar momentum exchanges have been considered between the fluid and discrete 27 element models (Flekkøy et al., 2002, Johnsen et al., 2006). These models have been shown 28 to closely reproduce experimental results, for the deformation of both granular materials 29 saturated with compressible fluids (Vinningland, 2007a, 2007b, 2010) and incompressible 30 ones (Niebling, 2010a, 2010b, 2012a, 2012b). In the present study, we will explicitly take 31 into account the momentum exchange (drag) between the fluid and the deformable solid, 32 considering the finite compressibility and the viscosity of the fluid. An important difference, 33 enabling us to address large systems, is that the solid will be treated here as a continuum with 34 a fault gouge rheology.

2 In this study, we build a conceptual simulation model to take into account elastic and 3 plastic porosity changes (e.g., Segall and Rice, 1995) and fault width evolution (e.g., 4 Yamashita, 1999), assuming that the fluid flow and seismicity expand predominantly along a 5 fault plane in a 3D medium. In particular, we address the issue of how induced seismicity, 6 once initiated on a fault, can be brought under control. As we aim to model micro-seismicity, 7 we treat the coseismic rupture process as simply as possible by solving the static equilibrium 8 equations of the elastic medium. In this respect, the approach for the pore pressure and solid 9 stress computations is similar to models developed for fluid injection by Rozhko (2010). Thus, 10 we are not introducing any coseismic thermal effects (Andrews, 2002), which are often 11 discussed for "large" natural earthquakes with large fault slip. The lubrication effect due to 12 pressurized fluid during unstable stages of the dynamics can also be taken into account 13 dynamically (Rice, 2006; Segall and Rice, 2006, Brantut et al, 2011), coupling the fluid 14 dynamics with a discrete elements model to represent the solid (Goren et al., 2011, Ghani et 15 al., 2013).

16 17

1

18 **2. The model** 

#### 19 **2.1 Conceptual model of a fault zone**

The geological structure and mechanical properties of the earthquake faults have been studied by field observations and drillings for certain active faults (e.g. Chester et al, 1993, Caine et al., 1996; Lockner et al., 2009). The fault core consists of: a rupture trace, surrounded by fault gouge, a damage zone and the surrounding host rock (Figure 1). The hydraulic properties also vary from the fault centre (core) to the host rock,

25 In this study, we will not be considering the poroelastic or the granular nature of the 26 fault core and damage zone. Let us assume that fluid is only allowed within the permeable 27 fault core (e.g., Segall and Rice, 1995; Rice, 2006). We also assume a variable finite width h28 of the fault core (Yamashita, 1999), as shown in Figure 1. We take the fault core thickness 29 into account when dealing with fluid behaviour, but we consider this thickness to be small 30 enough compared to the fault length for us to be able to calculate the elastic response of the 31 medium due to shear rupture. We then consider that the fluid only circulates in the fault core 32 and that the pore pressure reduces the effective normal stress applied on the fault (as observed at the scale of the fault thickness for impermeable lateral walls, see Goren et al., 2011, among others). Shear rupture is assumed to be described by the Coulomb criterion (e.g., Terzaghi, 1943). The rupture may change the state of the poroelasticity (e.g., porosity, permeability and fault width) of the fault core, but we assume that fluid flow and poroelastic response occur at very different time scales and can accordingly be solved sequentially (see numerical algorithm of Figure 2).

The system of the governing equations for fluid migration in any porous medium is
commonly based on two equations: firstly the linear Darcy's law if the Reynolds number is
sufficiently low (e.g., Chapman, 1981):

10 
$$\vec{q} = -\rho \frac{\kappa}{\eta} \nabla P$$
, (1)

11 which indicates that fluid mass flux  $\vec{q}$  [kg/s/m<sup>2</sup>] is proportional to the gradient of fluid 12 pressure with fluid density  $\rho$ , fluid viscosity  $\eta$  and permeability  $\kappa$ ; and, secondly, the 13 continuity of fluid mass:

14

$$\nabla \cdot \vec{q} + \frac{\partial (\rho \phi)}{\partial t} = \rho \dot{\Gamma}, \qquad (2)$$

- 15 where  $\phi$  is porosity and  $\dot{\Gamma}$  is the fluid source (volumetric injection rate).
- 16

20

One of the rheological models most frequently called upon is taken from Walder and
Nur (1994) and Segall and Rice (1995), who write the change in porosity as the sum of elastic
and plastic components:

$$\frac{d\phi}{dt} = \dot{\phi}_{elastic} + \dot{\phi}_{plastic} = \frac{\partial\phi_{elastic}}{\partial P} \frac{\partial P}{\partial t} + \dot{\phi}_{plastic} = \phi\beta_{\phi}\dot{P} + \dot{\phi}_{plastic}, \qquad (3)$$

21 where  $\beta_{\phi}$  is the elastic pore compressibility defined by  $\beta_{\phi} = (1/\phi)(\partial \phi / \partial P)$ . Strictly speaking, 22 the quantity  $\beta_{\phi}$  should represent the reversible poroelastic characteristic of the medium in the 23 case where  $\partial \phi_{plastic} / \partial P = 0$ . Combining the above equations, we obtain:

24 
$$\dot{P}(x, y, z) = \frac{1}{\phi(\beta_f + \beta_{\phi})} \left[ \nabla \left( \frac{\kappa}{\eta} \nabla P \right) - \dot{\phi}_{plastic} + \dot{\Gamma} \right], \tag{4}$$

corresponding to Equation (12) in Segall and Rice (1995), Equation (1a) in Wong *et al.*(1997), Equation (1) in Miller and Nur (2000) and Equation (7) in Goren et al. (2010).

#### **2.2.** Fault zone boundary condition and governing equation

In this study, we consider that the permeability  $\kappa$  varies along the (x, y)-fault plane as shown in Figure 1. We then integrate Equation (4) over the fault-perpendicular direction (z)for the fault zone as demonstrated in Yamashita (1999). We assume that the fluid does not flow across the interface between the fault core zone and the surrounding, low-permeability rock, at  $z = h_+$  and h., and that inside the fault core zone, the variables are uniform in the *z*direction, as variation along this direction is considered to be smaller than in the (x, y)directions:

9 
$$\int_{h^{-}}^{h^{+}} (\nabla \vec{q} + \dot{m}) dz = \int_{h^{-}}^{h^{+}} \rho \dot{\Gamma}(x, y, z) dz$$
(5)

10 which leads to:

11

18

$$h\left(\frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y}\right) + \left[q_z\right]_{h-}^{h+} + h\dot{m} = \rho h\dot{\Gamma}.$$
(6)

where the fault core width is expressed by  $h = (h_+) - (h_-)$ . Note that all the variables are hereafter averaged within the fault core and are a function of (x, y) only. There is no perpendicular flux across the fault boundary due to the impermeability of the surrounding medium, but it is related to the change of the boundary position itself, as follows:

16 
$$q_z = \rho \phi \frac{dh}{dt}$$
(7)

17 We then obtain:

$$h\left(\frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y}\right) + \rho \phi \frac{dh}{dt} + h \frac{d}{dt} (\rho \phi) = \rho h \dot{\Gamma} .$$
(8)

19 The product of the permeability  $\kappa$  and fault width h is called "transmissivity", which is often 20 used to describe the horizontal water flow in aquifers (e.g. Zimmermann and Bodvarsson, 21 1996; Zimmermann and Main, 2004). However, for our application, the shear rupturing may 22 greatly change the characteristics of the fault zone in terms of the porosity of the fault core 23 and the fault zone width. If either of these two parameters is unchanged during the process, 24 we could adopt the transmissivity as a model parameter. Now the equation is reduced to a 2D 25 problem (i.e., lubrication approximation). We can, therefore, similarly write, according to 26 Equation (4):

27 
$$\dot{P}(x,y) = \frac{1}{\phi(\beta_f + \beta_{\phi})} \left[ \nabla \left(\frac{\kappa}{\eta} \nabla P\right) - \phi \frac{\dot{h}}{h} - \dot{\phi}_{plastic} + \dot{\Gamma} \right]$$
(9)

28 This relation is to be compared with Equation (4) with an additional term including *h*.

The mass of the medium should be conserved regardless of the change in porosity:

2

1

 $\frac{d}{dt} \left( (1-\phi)\rho_{\phi}h \right) = 0,$ 

where  $\rho_{\phi}$  is the density of medium and then, again using  $\beta_{\phi} = (1/\phi)(\partial \phi / \partial P)$ , we obtain the 3 4 evolution for *h*:

5

$$\dot{h} = h \left( \frac{\dot{\phi}}{1 - \phi} - \beta_{\phi} \dot{P} \right) \tag{11}$$

(10)

(14)

Compared to the equation in Yamashita (1999), our boundary condition is defined differently 6 7 so that the conservation of mass is taken into account. Yamashita (1999) treated the fault core 8 width as an independent variable but in our formulation it depends on the other variables, 9 porosity and pressure.

10

#### 2.3. Rupture process and stress redistribution 11

12 As previously stated, the rupture process is governed by a Coulomb law. An increase 13 in pore pressure plays a role in reducing the fault strength. The fault strength  $\tau_f$  is expressed 14 as:

15

$$\tau_f = \mu_s \sigma_n^{eff} = \mu_s (\sigma_n - P), \qquad (12)$$

where  $\mu_s$  is the static frictional coefficient and  $\sigma_n^{eff}$  is termed effective normal stress, or 16 Terzaghi's normal stress (Terzaghi, 1943). The fact that this effective stress controls shear 17 18 rupture via a Coulomb law was experimentally established in triaxial laboratory tests, e.g., by 19 Nur and Byerlee (1971). The rupture does not begin if the applied shear stress is lower than 20 the strength. During the rupture, the fault strength reduces gradually with on-going slip ( $\Delta u$ ) 21 after a characteristic distance (so-called "slip-weakening distance" and usually called  $D_c$ ) to the residual stress (dynamic friction) level (Ida, 1972; Palmer and Rice, 1973; Ide an Takeo, 22 23 1997; Ohnaka, 2003; many others). This weakening process is often written as a simple 24 equation, such as:

$$\tau(\Delta u) = \tau_d + (\tau_f - \tau_d) \left( 1 - \frac{\Delta u}{D_c} \right) H \left( 1 - \frac{\Delta u}{D_c} \right), \tag{13}$$

where H(x) is the Heaviside function;  $H(x \ge 0) = 1$  otherwise 0. The residual strength 26

(dynamic stress)  $\tau_d$  is given through a dynamic friction coefficient  $\mu_d$  ( $\mu_d < \mu_s$ ): 27

28

25

29 The strength drop during an event is accordingly:

 $\tau_d = \mu_d \sigma_n^{eff} = \mu_d (\sigma_n - P)$ 

$$\Delta \tau = \tau_f - \tau_d = (\mu_s - \mu_d)(\sigma_n - P), \qquad (15)$$

if the effective stress does not change during this event. Comparing the fault weakening
process during an earthquake, the healing process is not well known. Usually, it is considered
that the fault heals over time and shear strength is progressively recovered (Dieterich, 1972).
However, the simplest, often used approximations, which are also the most extreme, are
immediate healing or no healing.

7

8 Equation (13) describes the relation between the on-going slip and strength evolution 9 on the fault. Slip generated on a fault element causes stress increases in the surrounding, so-10 called "stress redistribution". For simplicity, the stress redistribution is calculated as a static 11 dislocation problem in an infinite, homogeneous 3D elastic medium, a common procedure in 12 seismology. We regard the shear rupturing as occurring in a very narrow trace within the fault 13 zone. The static response function (Green's function) can be obtained analytically from the 14 equilibrium equation of elasticity. The discretization on square sub-faults is summarized in 15 Appendix 1 according to Tada et al. (2000). Tada et al. (2000) integrated the elastodynamic 16 equations by parts (renormalization process) so as to avoid the strong singularity when 17 estimating stress values along the fault, This stress redistribution may let the other fault 18 elements rupture subsequently. We iterate the process one element at a time until all the 19 elements are found to be stable (the imposed stress including the initial one and the 20 perturbation coming from the others is lower than the strength at that time). At last, we can 21 evaluate the seismic moment of one event through the traditional definition used in 22 seismology (e.g. Kanamori and Anderson, 1975):

23

$$M_0 = \mu \sum_i \Delta u_i \Delta s^2 , \qquad (16)$$

24 where  $\Delta u_i$  is the instantaneous fault slip at this time step for ruptured fault element *i*, and  $\Delta s^2$ 25 is the element's surface. One can follow the overall fault evolution with the cumulative fault 26 slip over time.

27

#### 28 **2.4. Fault porosity and permeability**

The link between porosity and permeability plays a crucial role. The permeability value depends strongly on the material and fracture network. Previously, Gentier *et al.* (2011a,

b), for example, found  $\kappa$  on an order of  $10^{-10}$  m<sup>2</sup> for a fracture network of the Soultz-sous-1 2 Forêt EGS site. Such a large effective permeability indicates that the network is relatively 3 well developed due to fractures and pre-existing fault networks. In laboratory experiments on porous materials, permeability is often about  $10^{-18}$  to  $10^{-12}$  m<sup>2</sup> (strongly dependent on the grain 4 5 size distribution; e.g. Kozeny, 1927; Carman, 1937; Løvoll et al., 2004; Johnsen et al., 2006; 6 Mavko et al., 2009), and thus the fault in the field must be highly fractured. However, the 7 microscopic rheology that links local porosity and permeability is quite complex, involving 8 plastic deformation processes that are difficult to describe (e.g. Bernarbé et al., 2003). The 9 relations between porosity and permeability are often power laws (e.g. Brace, 1977 and Wong et al. 1997), while they may change drastically from one state to another due to the rupturing 10 11 process (Miller and Nur, 2000). The permeability and the porosity describing the fault state 12 may change according to the fluid migration as well as shear rupturing (Figure 2). This is a 13 key point of the discussion in this study.

- 14
- 15

### 16 **3. Parameter studies**

#### 17 **3.1. Model setting and model parameters**

18 In order to understand the behaviour of our model, this section presents a parametric 19 study. The model parameters we employ are summarized in Table 1. We set an injection rate  $(\dot{\Gamma})$  of 31.5 [l/s] during the first 24 hours, for a fault plane consisting of 100 × 100 elements 20  $(3000 \text{ m} \times 3000 \text{ m})$ . Such an injection rate is quite common for EGSs during the stimulation 21 phases (Evans *et al*, 2012). The injection point, where the injection rate is  $\dot{\Gamma}$  is at the fault's 22 centre (everywhere else, the injection rate is nil). The element size is taken to be 30 m, which 23 24 limits the minimum size of possible earthquake in the simulation scheme. One may question 25 the resolution given by these parameters ; this is discussed in Appendix 2. The slip 26 weakening distance D<sub>c</sub> is taken as small enough so that this quantity is practically negligible for the used element size (see also in Appendix 2). At the model boundaries of the given fault 27 28 plane, we assume drained conditions.

#### **3.2. Finiteness of fault core width**

2 We first study some simple situations to understand the role of certain parameters. Let 3 us assume that the permeability is uniform and unchanging, regardless of the fault behaviour and fluid migration (Table 2). In Equation (3), we permit the elastic change of porosity 4  $(\dot{\phi}_{elastic})$ , but assume no plastic change  $(\dot{\phi}_{plastic} = 0)$ . The fault width is allowed to vary 5 following Equation (11). We test three different initial fault widths of  $h_0 = h|_{t=0} = 1, 3 \text{ and } 5 \text{ m}$ 6 7 as the models H1, H3, and H5 shown in Figure 3. We should bear in mind that the fault width 8 h behaves as a scale factor with respect to the permeability  $\kappa$ . In this configuration, fluid 9 behaviour is not linked with fault shear rupture (see also a snapshot at time  $t = 80\ 000\ s$ ). 10 Firstly, based on the assumed injection, pore pressure continues to increase at the fault's 11 centre. Then it decreases to zero after the end of the injection according to the diffusion term 12 of Equation (9) and a drained condition at the model boundaries. From a physical point of 13 view, pore pressure cannot be allowed to increase above a certain level. The upper limit, 14 therefore, is set at 40 MPa in these simulations in order to guarantee that the effective stress 15 remains positive. Physically, this represents a hydraulic fracture mechanism, where the confining walls of the fault fracture at this value limit the pressure at this saturation level. 16 Numerically, when P exceeds the upper limit, we impose  $\dot{P} = 0$ . Consequently, the system 17 18 absorbs the fluid mass change by increasing the fault width h. This type of feedback should 19 operate automatically in real cases. In model H1 ( $h_0 = 1$  m) in Figure 3(A), the pore pressure 20 immediately rises to the imposed upper value (i.e., when we do not apply this upper limit, the 21 pore pressure exceeds the confining pressure). A significant increase in fault width (of about 22 10 %) and porosity (300 %, also imposed as an upper limit) are necessary to absorb the 23 injected fluid mass, while those changes are negligible for the other two cases.

24

25 Seismicity is triggered by an increase in pore pressure; that is, the peak shear stress 26 required for rupture decreases versus the injection in model H3 ( $h_0 = 3$  m) in Figure 3(B). But 27 soon the shear stress increase soon comes to play a role in subsequent rupturing. In the shown example, the main swarm of seismicity ends with the largest event of  $M_w \sim 4.3$  (ruptured 28 dimension is 7.7 km<sup>2</sup>, or 8558 elements). After the largest event, moderate seismicity 29 30 continues and then decays slowly. The final event occurs half an hour after the injection has stopped in the case of  $h_0 = 3$  m. Several earthquakes have a magnitude of  $M_w \sim 1.4$ , 31 32 corresponding to the imposed minimum possible dimension of the potentially ruptured fault segment in the model, set to  $\Delta s = 30$  m. 33

2 When the fault width is large enough (model H5:  $h_0 = 5$  m in Figure 3(C)), we observe 3 that the fluid circulates rapidly without any significant pore pressure increase. Thus, 4 significant seismicity is not induced by the fluid injection. Achieving this condition is the 5 objective of industry-related fluid circulation. However, since we are interested in how the 6 seismicity evolves in our formulated system, we adopt the case of  $h_0 = 3$  m as our reference 7 for discussing the effects of various parameters and mechanics. We note here for our later discussion that the permeability  $\kappa$  is required to be an order of  $10^{-13}$  m<sup>2</sup> for a fault width of 8 h = 3 m for the shear rupturing to occur without increasing the pore pressure by too much 9 10 (Table 2).

11

1

#### 12 **3.3. Permeability evolution**

13 The fault property (porosity and permeability) evolves with time following the rupturing. In Equation (3), the first term originating from the elastic change of the porosity 14  $(\dot{\phi}_{elastic})$  is practically negligible, because it is estimated as of an order of  $10^{-4}$  [s<sup>-1</sup>] for a 15 porosity of 0.1 and a numerically possible change in pressure ( $\dot{P}$ ) of 10<sup>7</sup> [MPa/s]. As seen in 16 17 the later simulations, pressure changes are never so rapid in the considered cases. Therefore, the second term  $(\dot{\phi}_{plastic})$  may play a practical role, but the definition of  $\dot{\phi}_{plastic}$  still remains 18 19 quite uncertain. Segall and Rice (1995) introduced a relation that is analogous to the state 20 variable evolution of the rate- and state-dependent friction law. Yamashita (1999) describes 21 the plastic porosity as slip ( $\Delta u$ )-dependent:

22

$$\phi_{plastic} = \phi_{ss} + \Delta \phi \exp(-\Delta u / u_c) \tag{17}$$

where  $\phi_{ss}$ ,  $\Delta \phi$  and  $u_c$  are constant. Besides, the relation between the permeability and the porosity may be written as (e.g. Brace, 1977):

25  $\kappa = \kappa_0 (\phi / \phi_0)^n \tag{18}$ 

where  $\kappa_0$ ,  $\phi_0$  and *n* are constant. We tried to use these relations (Appendix 3). However the permeability does not always increase significantly and thus the pore pressure tends to increase easily in Equation (9), namely the insufficient increase in the porosity (the terms  $\nabla((\kappa/\eta)\nabla P)$  and  $\dot{\phi}_{plastic}$ ) does not always cancel the injection term ( $\dot{\Gamma}$ ). In other words, 1 alternative evolution of  $\kappa$  is necessary due to the external conditions (rupturing, stress, etc.) 2 rather than the internal relations, such as Equations (17) and (18).

Therefore, we adopt the toggle switch (e.g. Miller and Nur, 2000) in which the permeability changes independently from the change in porosity. For simplicity, we introduce two irreversible phases locally according to the state of the fault at each point, i.e., intact or already ruptured:

7 
$$\kappa(x, y) = \begin{cases} 10^{-14} m^2 \text{ for } u(x, y) = 0\\ 10^{-13} m^2 \text{ for } u(x, y) \neq 0 \end{cases}$$
(19)

where *u* is the cumulative slip ( $u = \sum \Delta u$ ). Thus, the value is initially small everywhere, and 8 once rupture has occurred, permeability increases up to a constant value of  $10^{-13}$  m<sup>2</sup>. The 9 value of 10<sup>-14</sup> m<sup>2</sup> assigned to the material before shear rupture took place might be much 10 smaller as in granular materials; however, the simulation is not influenced, as the given value 11 12 is quite small so that the fault is practically impermeable at the time scale of our interest. No plastic change in porosity is assumed ( $\dot{\phi}_{plastic} = 0$ ), as the permeability is no longer related to 13 the porosity. The term  $\nabla((\kappa/\eta)\nabla P)$  is dominant in Equation (9). We show the simulation 14 15 results (hereafter referred as model K2) in Figure 4. The other parameters are the same as for 16 model H3 in Figure 3(B). As the pore pressure increases along the boundary of permeability 17 contrast, that is, at the front of the rupture zone, seismicity migrates away from the injection 18 point. Seismicity is far more pronounced than model H3 in Figure 3(B): the rupture front 19 advances gradually, step by step. The pore pressure does not increase beyond the rupture front 20 due to the low permeability, so that it is difficult for rupture to occur there, while in the 21 previous case, the pore pressure increases over a wide area. The assumption of such an abrupt 22 change in permeability due to fracturing leads to seismicity being caused by fluid migration as 23 the injection continues.

- 24
- 25

#### **3.4. Heterogeneity in stress field and fault strength**

The origin of the heterogeneity observed in earthquake dynamics (e.g. Kanamori and Stewart, 1978; Aki, 1979) is a fundamental question in seismology. Schmittbuhl *et al.*, (2006) study the stress field of the Nojima fault before and after the 1995 Kobe, Japan, earthquake and propose that the fluctuations of the stress field along the fault would be dominated by "quenched" fault properties rather than dynamic stress fluctuations produced during the earthquake. Let us study the effect of heterogeneity in initial shear stress or in fault strength on the produced seismicity. The meaning of heterogeneity differs for the stress field and the fault strength, since the former can be released by a rupture while the latter may remain indefinitely despite repeated ruptures. We first furnish a heterogeneous shear stress, simply assuming a periodic variation according to a single Fourier mode approach as a first step, although the actual stress field could be a superposition of different characteristic lengths of heterogeneity (e.g., Mai and Beroza, 2002):

$$\tau_0 = T_0 \times \left[ 1 - \delta \cdot f(x, y) \right] \tag{20}$$

where  $f(x, y) = \sin(2\pi x / \lambda_x + \alpha_x) \sin(2\pi y / \lambda_y + \alpha_y)$  is a function of position (x, y),  $\lambda_x$  and  $\lambda_y$ 9 are the given wavelengths in spatial heterogeneity, and  $\alpha_x$  and  $\alpha_y$  are random numbers. The 10 11 variable  $\delta$  indicates the amplitude of the heterogeneity, namely a ratio with respect to the 12 absolute amplitude. In Figure 5, we show the simulation results for (A) model K2\_ST10 with  $\delta$  of 10 % ( $\delta$ = 0.1) and (B) model K2\_ST20 with  $\delta$ = 0.2, taking  $\lambda_x$  = 300 m and  $\lambda_y$  = 500 m, 13 respectively. Introducing the heterogeneity leads to more overpressure locally and yields more 14 15 earthquakes comparing to model K2 of Figure 4. The snapshots are also shown for model 16 K2 ST20 with  $\delta = 0.2$  (i.e., 20 %) at different times (Figure 5). The expansion of the ruptured area is heterogeneous and complex in space. In snapshot (1), we note that the area of slip 17 18 extends slightly further along the y-axis because of the longer wavelength of the 19 heterogeneity. For instance, Perfettini et al. (2001) numerically study the correlation of the 20 slip pattern and strength heterogeneity in seismic cycles and find that the slip appearance reflects the background strength heterogeneity. This is also the case in our simulation. When 21 22 the fault has almost entirely ruptured, the fluid circulation becomes quasi-uniform because of 23 the homogeneous permeability in the ruptured area. At this time, some areas remain 24 unruptured (snapshot at time (2) in Figure 5).

Similarly, we study the effects of spatial heterogeneity for the other parameters. The fault strength, i.e., the static frictional coefficient, is given a variation based on the reference frictional coefficient  $\mu_{s0}$ :

28

$$\mu_{s}(x, y) = \mu_{s0} \times \{1 + var \cdot (\mu_{s0} - \mu_{d}) f(x, y)\}$$
(21)

where the same variation function f(x,y) as in Equation (20) is used. We note that normal stress itself does not change since we are considering a planar fault, but the effective normal stress changes due to the pore pressure change (Equation (12)). The dynamic frictional 1 coefficient  $\mu_d$  is assumed to be uniform everywhere. The relation in Equation (21) represents a 2 continuous change, but we also assume heterogeneity in the discontinuously localised forms:

3 
$$\mu_{s}(x, y) = \begin{cases} \mu_{s0} \times \{1 + var \cdot (\mu_{s0} - \mu_{d})\} & \text{if } f(x, y) \ge 0.8 \\ \mu_{s0} & \text{otherwise} \end{cases}$$
(22)

4 Fault strength after the first rupture is always heterogeneous according to Equations (21) and 5 (22). We show, in Figure 6, the snapshots for two cases, (A) model K2\_TC20 using Equations (21) and (B) model K2\_TD20 using Equation (22), with  $\delta = 0.5$  (50 %), that is, the 6 7 maximum value of a given  $\mu_s$  is 0.975. Slightly more earthquakes are observed in model 8 K2\_TD20 (B) than model K2\_TC20 (A) from the histograms. Behind the rupture front, in 9 model K2 TD20 (B), the heterogeneity remains visible and the overall fault slip is smaller 10 when comparing the two snapshots at  $t = 16\ 000\ s$ . From these simulations, we discover that 11 discontinuous localised heterogeneity affects the details of the rupture growth. The rupture 12 front shape becomes very heterogeneous, and heterogeneous slip distribution persists behind 13 the rupture front. As a result, the total slip is reduced.

14

#### 15 **3.5. The appearance of seismicity**

In Figure 7, we analyse the magnitude-frequency relation from some of the simulated seismicity catalogues. In all cases, we obtain the classic power law relation known as Gutenberg-Richter (Gutenberg and Richter, 1954) with a slope of about -1. Large-magnitude events have only a single sample for a given magnitude range of 0.1, and they seem to diverge from the power law. However, the ruptured area for large events reaches the model boundary, so this point should be regarded with caution.

22

23 In general, the number of earthquakes increases at the same time as larger-magnitude 24 earthquakes appear. Pore pressure increases until a large earthquake (Mw > 4) occurs, 25 creating a continuous fluid path to the drained model boundary, i.e., a breakthrough. Once this 26 happens, the system becomes stable, and the level of seismicity reduces. This silence is 27 somehow an artificial effect, as the size of the maximum event and the ruptured area is limited 28 by the *a priori* model dimension. However, seismic hazard assessment for induced seismicity 29 in geothermal sites is a major concern (e.g., Majer et al., 2007; Baisch et al., 2009), although 30 phenomena are not completely understandable mechanically (Majer et al., 2012).

### **4. The potential for seismicity control**

2

#### 4.1. The effect of injection and spontaneous seismicity propagation

A challenging question is to find out how the seismicity can be controlled by the injection. We reduced the injection rate by ten compared with the previous cases and its duration to a few hours instead of one day, expecting to observe an evolution in seismicity (or no seismicity) once injection had ceased.

7

8 As shown in Figure 8, we simulate two cases under the homogeneous condition (A: 9 model K2\_I1) similar to model K2 in Figure 4 and heterogeneous initial shear stress (B: 10 model K2\_ST20\_I2) like model K2\_ST20 with  $\delta = 0.2$  (20%) of Figure 5. The onset of 11 induced seismicity is delayed simply because it takes much more time for sufficient pore 12 pressure to build up at the injection point, while the duration of seismicity does not change 13 visibly as compared to previous cases. The duration of the injection is three and four hours 14 respectively. The difference in injection duration comes simply from the fact that the given 15 heterogeneity (model K2 ST20 I2) needs a greater concentration of pore pressure at the 16 injection point. In both cases, the injection is stopped just after the seismicity began. However, 17 the seismicity continues to propagate and eventually ruptures the entire model fault. We also 18 notice that introducing heterogeneity does not affect this rupture expansion, although it does 19 generate much more seismicity and pore pressure remains relatively high. The rupture 20 progress is locally somehow blocked by the heterogeneity but continues on to the end when 21 the fault system is relaxed.

22

23 We conclude that it is very difficult to control seismicity through injection under the 24 given conditions. Once induced, the seismicity propagates rather spontaneously, probably 25 because the heterogeneity considered in this study might still be too small in amplitude and 26 too short in its characteristic correlation length of heterogeneity (Perfettini et al., 2001). The 27 expansion rate of seismicity is insensitive to differences in injection protocol. The shear 28 rupture spontaneously triggers other ruptures to the surrounding area. In this sense, the 29 seismicity is self-induced once started. The mechanism might be similar to the propagation of 30 solitary dislocation propagation (Schmittbuhl et al, 1993), or to Bürgers-like solitons 31 evidenced theoretically for the propagation of overpressure pulses in mud volcanoes (Garcia *et al.* 2000, Revil 2002). The source of this mechanism lies in the nonlinear diffusion process
 for the fluid pressure, arising due to the pressure dependence of the permeability.

3

#### 4 **4.2.** Fault healing and trapped fluid circulation

5 The real permeability of the fault should be more complex than given by Equation (19). 6 We observe that in all the previous simulations, pore pressure finally drops to zero once 7 injection stops. All the injected fluid flowed out from the model boundary as the fault became 8 sufficiently permeable. In reality, the permeability may decrease over time to disturb the fluid 9 circulation and trap it. Let us, therefore, introduce the temporal evolution (sealing) of 10 permeability relative to fault healing, analogous to Aochi and Matsu'ura (2002):

11 
$$\frac{d\kappa}{dt} = -\beta(\kappa(t) - \kappa_{\infty})$$
(23)

where the parameter 1 /  $\beta$  provides the characteristic time so that the permeability converges 12 13 to the final value of  $\mathcal{K}_{\infty}$ . We do not seek to determine here whether the mechanism of the 14 healing process is mechanical or chemical. For the sake of simplicity, we assume  $\kappa_{\infty} = \kappa(t=0)$ ; however, the fault core may take on a new value, different from the initial 15 permeability, because of the development of a micro-fracture network and the shear rupture. 16 No healing takes place in the limit of  $\beta = 0$ , and  $\beta = 1/(432,000 \text{ s}) = 1/(5 \text{ days})$  does not 17 produce any visible effect on the simulation. On the other hand, a large  $\beta$  value indicates a 18 19 rapid healing of the permeability. For example, in the case of  $\beta = 1/(60 \text{ s})$ , the healing is 20 exceedingly fast, and pore pressure becomes very high at the injection point. In our current 21 system, the fluid circulation is allowed, not during the coseismic event, but after this rupture. 22 However, the permeability becomes low again soon after the rupture. Thus the fluid 23 circulation is limited. If we allow fluid migration during the event, say  $\kappa = \infty$  on the 24 concerned fault segments during rupture (Miller and Nur, 2000), it would be possible to 25 introduce an immediate healing process.

26

In Figure 9, we show a typical intermediate case of  $\beta = 1/(21\ 600\ s) = 1/(6\ hours)$ , model K2\_B6, under the same conditions as model K2 in Figure 4. After the first swarm of seismicity terminated by rupture of the entire fault (the first six hours), there is a period of quiescence. This quiet period allows time for the fault to heal, specifically because of the decrease in permeability in the fault zone. As a result, the fluid starts to be trapped, and since

1 fluid is continuously injected, seismic activity resumes once again releasing fluid. After the 2 end of the injection, the fluid is not completely drained due to this sealing effect. Thus, the 3 pore pressure remains at a certain finite value. In a natural context, such distinct swarms may 4 not be identified, as the system can be expected to be more complex and heterogeneous than 5 the simplified model (a single, uniform parameter of sealing). The fact that the system transits 6 from a continuous activity to an intermittent one when  $\beta$  increases is the analogue, in induced 7 seismicity, to the transition observed for fault activities under tectonic loading, when the 8 healing process changes from slow to fast rates - as was observed and explained along the 9 San Andreas Fault (Gratier, 2011, Gratier et al., 2011).

10

11 We now try to emphasize the healing effect on the seismicity and fluid circulation, by 12 reducing the injection time and assuming  $\beta = 1/(3\ 600\ s) = 1/(1\ hour)$ . We test homogeneous and heterogeneous initial shear stress ( $\delta = 0.2$ ) with wavelengths of  $\lambda_x = \lambda_y = 300$  m, 13 respectively (A: model K2\_B1, and B: model K2\_SH20\_B1 in Figure 10). This version of the 14 15 model reproduces the seismicity without fracturing the whole model fault and the trapped 16 fluid effect without breakthrough. The pore pressure after the end of injection decreases not to 17 zero but to a finite value. Unlike the previous simulations, we also find that heterogeneity 18 plays an important role. In model K2\_B1 in Figure 10(A), homogeneous initial shear stress 19 (the same as for model K2 in Figure 4 except for  $\beta$ ), seismicity gradually but rapidly ceases 20 after the end of the injection (1 hour). In model K2\_SH20\_B1 (Figure 10(B)) employing 21 heterogeneous initial shear stress, it takes a much longer time for the seismicity to subside. 22 The seismicity rate (number per hour) significantly increases after the end of the injection, 23 and we observe an earthquake of magnitude 4.39 triggered close to the model boundary. This 24 significant perturbation can be due to the particular distribution of heterogeneity. However, 25 the seismicity ceases after approximately nine hours and fluid is ultimately trapped on the 26 healed fault with a relatively high pore pressure.

27

### **4.3 Fault interaction**

One of the major hypotheses in this study is that the rupture process is restricted to a single fault plane. In this case, the stress redistribution is carried out strictly following the equations in Appendix A, bringing stress disturbance at any distance. However, as previously treated in Baisch et al. (2010), let us assume that the stress disturbance only affects the 1 neighbouring elements, namely,  $G_{(i-l)(j-m)} = 0$  for |i-l| > 1 or |j-m| > 1 in Equation (A2). 2 This extreme condition may correspond to a segmentation of the fault off the major plane.

3

4 We show an example of seismicity evolution according to the given injection rate 5 (model SB in Figure 11). This time, rupture is significantly influenced by the injection 6 process despite the fact that the ruptures grows quickly independently of the injection rate. In 7 order to demonstrate how the seismicity is sensitive to the injection protocol, we assume a 8 synthetic injection protocol varying during 2.5 days. We also introduce a relation for the 9 permeability evolution such that it increases by 50 % with every rupture until the upper limit. 10 The seismicity increases during the course of the injection and reducing the injection rate 11 decreases the seismicity. When the injection rate increases once again the seismicity starts to 12 evolve. The seismicity continues during the whole duration of the injection and after its end.

13

The effect of stress redistribution is important in governing the induced seismicity. The structure of natural fault systems can be more complex rather than a single fault plane. It is worth of pointing out that stress redistribution on a single planar fault tends to accelerate the rupture growth rapidly, namely "self-induced" seismicity, independent of the injection protocol.

19

### 20 5. Discussion

21 A simple spring-block model does not take into account the elastic response at 22 distance, but only distributes the stress over neighbouring elements, so that the initially 23 imposed heterogeneity also generates a strong heterogeneity in stress and fault slip (e.g., Bak 24 and Tang, 1989). On the other hand, elastic systems which have remote responses (e.g., Rice, 25 1993) behave in such a way as to homogenize both stress and slip. The friction law assumed in this study does not consider any evolution process during a single coseismic event; that is, 26 27 it lacks a length scale such as characteristic slip distance. Thus, inherently, our discrete 28 system might display spatio-temporal complexity, but globally over a large spatial scale and a 29 long time scale, the simulation finishes when the fault has completely ruptured and/or a 30 stationary injection flow is established from the injection point to the outer boundary of the 31 model. All the simulations display spatial migration, with many events statistically

reproducing the scheme of the Gutenberg-Richter magnitude frequency relation. Once
 seismicity is induced, we see that it is difficult to control because the shear rupture process
 becomes dominant. This self-induced behaviour is different from any model simulated by a
 spring-block model of interaction with the immediate neighbourhood (Baisch *et al.*, 2010).

5

6 We summarize the spatio-temporal evolution of the seismicity simulated in this study 7 (Figure 12). The seismicity, once induced by the injection, clearly propagates rather 8 spontaneously, independently of the injected fluid in our simulations. Seismicity migration is 9 often represented in the form:

10

$$r = \sqrt{4\pi Dt} \tag{24}$$

where r is the distance of the seismicity front from the injection point, t is the time since the injection onset and D is hydraulic diffusivity (e.g., Shapiro and Dinske, 2009). Shapiro and Dinke (2009) also report that there are some local clouds presenting a very rapid linear trend of seismicity migration that may be related to the fracture opening and reopening. Conceptually, our single-fault core model corresponds to the latter situation.

16

17 In our simulated system, we find that the evolution of fault permeability is essential. 18 We have considered the hypothesis in which the permeability immediately increases after 19 fracturing and gradually decreases with time, given by the simple constitutive relations in 20 Equations (17) and (23). The form of these equations is still to be explored, as many 21 researchers propose different evolution laws and extreme values are possible (e.g., Miller and 22 Nur, 2000); this said, our idea is qualitatively consistent with earlier conceptions (e.g., Miller 23 and Nur, 2000). But permeability may increase even before or during rupture, and aseismic 24 slip is possible (Calò *et al.*, 2011). We have simplified our 3D system to a projected 2D fault 25 model using the hypothesis that fluid migration and fault movement occur dominantly within 26 a fault core. All of the injected fluid mass must circulate within the fault core along the fault 27 plane and no loss is allowed off plane. No extraction from other wells is considered but 28 further studies would be called for. In the field, the seismicity may not always be aligned on a 29 single plane but it may be clustered in a 3D volume. We also find that the pore pressure 30 sometimes becomes extremely high when the given injection rate, permeability, porosity and 31 fault width are not suitable. However, in a real system there should be some mechanical 32 feedbacks between these parameters as well as an off-plane loss mechanism to keep pore 33 pressure reasonable at any given time.

### 3 6. Conclusion

4 We developed a system of equations describing fluid migration, fault rheology and 5 shear rupture. Assuming that the phenomena predominantly take place on a single fault with a 6 finite permeable zone of variable width, we are able to project the volumetric fault core 7 equations onto the 2D fault plane by introducing a redefined boundary condition of Equation 8 (10). Then we propose a toggle-switch type evolution of fault core permeability, Equation 9 (19), and a healing process, Equation (23). Several parametric studies are performed to 10 understand the basic behaviour of the system established by injecting the fluid at a single 11 point. Fault rheology is a key element. In the absence of fault healing (no decrease in 12 permeability) after rupture, seismicity is generally self-induced, e.g. tends to evolve 13 spontaneously once it is triggered, independently of the injection, and the fluid is completely 14 drained. However, when a sufficiently rapid healing process takes place, the fluid mass is 15 locally trapped along the fault and rupture migration can occur repeatedly. The pore pressure 16 remains high after the end of the injection. The effect of segmentation of stress redistribution 17 in the simulation has been studied and, in such cases, the system is sensitive to the injection protocol. 18

19

In this study, we have not precisely calibrated each parameter for specific examples of induced seismicity. The model developed in this study is general for different aspects of seismicity, such as plate boundaries, which may involve large amounts of fluid, as well as for earthquakes induced following injection or extraction of fluid or gas into or out of reservoirs.

24

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5

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### **Appendix 1: Elastic equation for stress redistribution**

11 The theory of the linear static elasticity gives us a unique solution based on the 12 equilibrium theory. The stress change at any place in the medium is written as a convolution 13 of the causal fault slip over and the response function (Green function), *G*, *over* the whole 14 area of the fault ( $\Sigma$ ):

$$\Delta \tau(\vec{x}) = \int_{\Sigma} G(\vec{x} - \vec{\xi}) \Delta u(\vec{\xi}) d\Sigma.$$
 (A1)

16 The response function *G* has symmetry with respect to the relative location  $(\vec{x} - \vec{\xi})$ . Let us 17 discuss only on a fault plane (z = 0) and discretize (x, y)-fault plane in squares of the equal 18 size;  $\vec{x} = (i\Delta s, j\Delta s, 0)$  and  $\vec{\xi} = (l\Delta s, m\Delta s, 0)$ . Then we can write (A1) as

19 
$$\Delta \tau_{ij} = \sum_{l,m} G_{(i-l)(j-m)} \Delta u_{lm} .$$
 (A2)

Tada et al. (2000) provide the analytical solution for the 3D homogeneous, elastic, isotropic
medium. For a dislocation in the *x*-direction, shear stress in *xz*-component is written as

22 
$$\Delta \tau(\vec{x}) = -\frac{\mu}{4\pi} \int_{\Sigma} d\Sigma \left[ 2(1-p^2) \frac{\gamma_1}{r^2} \frac{\partial \Delta u}{\partial x} + \frac{\gamma_2}{r^2} \frac{\partial \Delta u}{\partial y} \right]$$
(A3)

where  $p^2 = \beta^2 / \alpha^2$  ( $\alpha$ : P-wave velocity,  $\beta$ : S-wave velocity),  $r^2 = \left| \vec{x} - \vec{\xi} \right|^2$  and  $\gamma_i = (x_i - \xi_i) / r$ . In such boundary integral equation formulations, slip on an element is usually assumed uniform;

$$\Delta u(\bar{\xi}) = \Delta u_{lm} \Big[ H(\xi_1 - (l - 1/2)\Delta s) H(\xi_2 - (m - 1/2)\Delta s) - H(\xi_1 - (l - 1/2)\Delta s) H(\xi_2 - (m + 1/2)\Delta s) - H(\xi_1 - (l + 1/2)\Delta s) H(\xi_2 - (m - 1/2)\Delta s) + H(\xi_1 - (l + 1/2)\Delta s) H(\xi_2 - (m + 1/2)\Delta s) \Big]$$
(A4)

2 and we estimate the stress at the center of each element  $(i\Delta s, j\Delta s)$ . We can obtain the 3 analytical solution from (A3) for a unit slip  $\Delta u = \Delta u \cdot H(x_1)H(x_2)$ ;

4 
$$\Delta \tau(\vec{x}) = -\frac{\mu \Delta u}{4\pi} \left[ \frac{1}{\sqrt{x_1^2 + x_2^2}} \left\{ 2(1-p^2) \frac{x_2}{x_1} + \frac{x_1}{x_2} \right\} - \left\{ 2(1-p^2) \frac{1}{x_1} + \frac{1}{x_2} \right\} \right].$$
(A5)

Consequently, we can obtain the response function for Equation (A4) by combining the
solution (A5), as illustrated in Figure A1. Note that the stress remains finite at any collocation
points, including on the causal source element (0, 0) and the neighbouring elements (±1, ±1).
The detailed derivation of the equations is given in Tada et al. (2000).

9

### **Appendix 2: Resolution in numerical simulations**

11 In the overall study, we suppose an element size of 30 m, which allows a minimum 12 earthquake of about M1.3 under the given friction law, namely, Equation (16). As the system 13 of the equations is a highly nonlinear self-organising system, we are concerned about the 14 resolution of the numerical simulations seen at different scales. Figure A2 represents a 15 simulation with a grid size of 50 m (time step of 2 s) for the same physical parameters used in 16 Figure 4. As expected the minimum magnitude appearing during the simulation increases to 17 about 1.8, namely smaller earthquakes than this are not modelled. Consequently, the number 18 of earthquakes decreases. Due to the nonlinearlity and different discretization, we may not 19 find the same earthquake (epicentre location, magnitude, time, etc.) in a deterministic manner, 20 strictly speaking. However the statistical features of the seismicity is the same. The seismicity 21 propagates outwards and then we find a maximum event of Mw 4.47 at 4.9 hours after the 22 injection start (previously Mw 4.52 at 4.1 hour in Figure 4). Most of the seismicity is calmed 23 after 6 hours. The pore pressures on the fault (lines in red and orange) show an identical 24 behaviour at different points. These observations assure the coherence in our discussions on 25 the statistical characters of the seismicity evolution and fluid circulation.

26 On the other hand, one may also pose a question about the slip-weakening distance 27 (*Dc*) used in this study. We dare to take it very small, because we do not want to bring any 1 complex discussion about its temporal evolution (e.g. Nakatani 1997; Aochi and Matsu'ura, 2 2002) or its scaling problem (e.g. Ohnaka, 2003; Aochi and Ide, 2009). However it is worth 3 mentioning what may happen if we take a longer Dc. The fact of small Dc is that the rupture 4 brings the totality of the stress drop described by Equation (15) regardless of the dimension of 5 rupture (number of ruptured elements). This is why the smallest earthquakes are aligned 6 around a magnitude 1.3. Later on, smaller earthquakes appear, because an increase in the pore 7 pressure leads to a decrease in effective normal stress and consequently a smaller stress drop.

8 Figure A3 represents a simulation assuming Dc = 0.01 [m] under the same 9 configuration with Figure 4. We note that Dc = 0.01 [m] is a reasonable value for an event of magnitude 4 (Ohnaka 2003; Ide and Aochi, 2005). A long Dc allows a tiny displacement with 10 11 a smaller stress drop (a very small magnitude event), and this generates only a slight stress 12 concentration in the surrounding (schematic illustration in Figure A4). In order to accumulate 13 enough shear stress in the surrounding, many small earthquakes are required. On the other 14 hand, as described in Equation (19), the permeability becomes immediately large once the 15 concerned element is ruptured regardless of its slip amount. Thus, the fluid circulation may 16 later on play a role in reducing the fault strength during the seismicity. As a result, the 17 seismicity appearance becomes very complex, and their magnitude-frequency relation is not 18 any more a continuous linear inverse relation. Taking into account a finite length of Dc is 19 required particularly when considering a certain pre-fixed size of earthquake and its 20 preparation process (e.g. towards a M4 event in this case), however this does not assure the 21 self-organizing system of multi-scale phenomena. For this purpose, ones will have to carry 22 out dynamic rupture simulations taking into account of the inertia and scaling problem (e.g. 23 Aochi and Ide, 2009). However for the purpose of this study and for the simplicity of the later 24 discussion, we adopt the assumption that Dc is small enough to let the complete stress drop 25 every time.

26

### **Appendix 3: Preliminary attempt on fault evolution**

As our preliminary attempt, we have tested the continuous, relatively gentile, relations, Equations (17) and (18). In Figure A5, we show two test cases in the first twelve hours during the continuous injection, supposing  $u_c = 0.01$  m in Equation (17) and n = 5 or 10 in Equation (18). The porosity and the permeability are shown for the central element where the fluid is

1 injected. In both cases, the system finds rapidly the equilibrium status (fluid circulates) after 2 only a few events without any significant seismicity. As observed in the lower panels, the 3 permeability and the porosity changes are so quick, and practically discontinuous, and are 4 also saturated, as the porosity is forced to be limited to 0.45, although this sounds extremely 5 large. As the rupture process is nonlinear, the slip amount of successive events differ from 6 each other are very difficult to control, while the relations such as Equations (17) and (18) are 7 unique. These examples show the difficulty to control the fluid circulation in simple equations. 8 Thus we will adopt a discontinuous phase change in permeability change according to the 9 rupturing.

Table 1: Model parameters 

Parameter	Quantity
Element size $\Delta s$	30 m
Time step $\Delta t$	1 s
Fluid viscosity $\eta$	$2 \times 10^{-4}$ Pa.s
Fluid density $\rho$	$1 \ge 10^3 \text{ kg/m}^3$
Fluid compressibility $\beta_{f\phi}$	$5 \ge 10^{-10} \text{Pa}^{-1}$
Elastic medium compressibility $\beta_{\varphi}$	$5 \ge 10^{-11} \text{ Pa}^{-1}$
Injection rate $\Gamma$	31.5 [l/s] during the first 24 hours
Rigidity of elastic medium $\mu$	30 GPa
Normal stress (confining pressure) $\sigma_n$	100 MPa
Background pore pressure	30 MPa
Static friction coefficient $\mu_f$ , $\mu_f'$	0.65 (for the first rupture), 0.6 (for the rest)
Dynamic friction coefficient $\mu_d$	0.55
Critical slip displacement $D_c$	< 0.001 cm

Table 2: Three fault parameters in the first examples.

	-	
Parameter	Initial value	Evolution allowed
Fault width <i>h</i>	1, 3 or 5 m	Variable
Permeability $\kappa$	$10^{-13} \text{ m}^2$	Invariable, const
Porosity ø	0.05	Variable only by elastic change

1 Figure captions

Figure 1: Conceptual illustration of fault internal structure (after Chester et al., 1994) and our
model allowing fluid circulation and induced seismicity predominantly along a
preexisting 2D fault plane with a fault core of width h(x, y). The fault core is
surrounded by impermeable rock. Fault slip is given by Δu(x, y) directed along the xaxis, so that the shear stress τ(x, y) of interest is a xz-component. The pore pressure in
the fault core P(x, y) can evolve due to the injection Γ(x, y), as a function of the
variable fault constitutive parameters of porosity φ(x, y) and permeability κ(x, y).

Figure 2: Numerical algorithm used in this study. In the left hand side, the loop is concerned
about the fluid circulation. Once the rupture criterion is reached (left bottom), the
rupture process is solved at the right hand side. Every time, fault property (porosity and
permeability) may evolve.

Figure 3: Simulated seismicity versus time in the case of invariable permeability ( $\kappa = 10^{-13}$ 13 14  $m^2$ ) for different initial fault widths ( $h_0$ ) of 1, 3 and 5 m, in (A) model H1, (B) model H3 15 and (C) model H5, respectively. The seismicity is plotted by dots, colored to reflect the 16 epicentral distance from the injection point, and the number of earthquakes per hour is 17 also shown in the form of histograms. The injection rate is indicated by blue lines. The 18 pore pressure, represented by red and orange curves, corresponds to two different spots, 19 the injection point (x, y) = (1500 m, 1500 m) and another point (1290 m, 1290 m), 20 respectively; their position is indicated by triangles in a snapshot (cumulative slip, pore 21 pressure, permeability and shear stress) at time t = 80,000 sec, indicated by a line (1).

- Figure 4: Simulated seismicity with variable permeability. Hereafter it is referred as model K2.
  Snapshots are shown at two different times, labeled (1) and (2). See also the caption for
  Figure 3.
- Figure 5: Simulated seismicity for two heterogeneous fields of initial shear stress, (A) model K2\_SH10:  $\delta = 0.1$  (i.e., 10%) and (B) model K2\_SH20: 0.2 (20%). Snapshots are shown at two different times, labeled (1) and (2). See also the caption for Figure 3.
- Figure 6: Simulated seismicity for two descriptions of heterogeneous fault strengths, (A)
  model K2\_TC20: spatially continuous and (B) model K2\_TD20: discontinuous cases.
  Snapshots are shown at two different times labeled (1) and (2). See also the caption for
  Figure 3.
- Figure 7: The magnitude-frequency relations for different simulations. The first panel shows
   four different cases of heterogeneous initial shear stress field, model K2 (Figure 3),

- 1 K2\_ST10 (Figure 4(A)), K2\_ST20 (Figure 4(B)) and K2\_ST30 with  $\delta = 0.3$  (30 %). The 2 second panel shows the cases of different wavelengths based on K2\_ST20.
- Figure 8: Simulated seismicities for short injection durations. (A) Model K2\_I3:three hours of
  injection for the homogeneous condition (the same configuration as model K2 in Figure
  4), and (B) model K2\_ST20\_I4: four hours of injection for heterogeneous condition (cf.
  model K2\_ST20 in Figure 5). The latter case requires a longer injection because the
  initial shear stress at the injection point happens to be low.
- 8 Figure 9: Simulated seismicity for the case of β = 1/(21,600 s), model K2\_B6, for an
  9 injection of 24 hours (the same as Figure 4 except for β).
- Figure 10: Simulation results for the cases of  $\beta = 1/(3600 \text{ s})$  for a short injection duration. (A) Model K2\_B1: homogeneous and (B) model K2\_SH20\_B1: heterogeneous initial shear stresses are given, respectively, letting  $\lambda_x = \lambda_y = 300 \text{ m}$ . Snapshots are shown for the two different moments for each simulation. In the latter case, the pore pressure is trapped after injection has stopped and the fault is not entirely ruptured.
- Figure 11: Evolution of seismicity according to varied injection rate with time (model SB).
  The stress redistribution affects only the adjacent elements and the permeability changes
  evolutionally every event.
- Figure 12: Evolution of seismicity for the four cases in time-versus-distance graphs, models
  K2 (Figure 4), K2\_SH20\_I4 (Figure 7(B)), K2\_B6 (Figure 9) and K2\_SH20\_B1 (Figure 10(B)). The injection rate is shown by grey lines. The dots (individual earthquakes) are color-coded for magnitude.
- Figure A1: Shear stress redistribution in the 3D homogeneous elastic medium due to a
  dislocation (slip direction is parallel to the x-axis) located at the center of the fault plane.
  All the axes are normalized.
- Figure A2: The same situation as model K2 in Figure 4 but with an element size of 50 m for checking the numerical resolution. The statistical feature of the appearing seismicity and the fluid circulation are consistent in the both simulations. For notations, see also the caption for Figure 3.
- Figure A4: The same situation as model K2 in Figure 4 but with a critical slip displacement *Dc* of 0.01 m.
- Figure A3: A schematic illustration of the stress release and redistribution in the cases of (1)
  immediate stress drop (*Dc* small enough) and (2) a longer *Dc*. As illustrated at the top
- 33 corner, the slope (grey line) of the relaxation is determined by the system of the elastic

1 equations. The discrepancy between the charged shear stress and the fault strength is 2 equilibrated by an increment slip. Namely the equilibrium state is the cross between the 3 relaxation line (solid grey) and the friction lines (dotted or broken lines). If Dc is small 4 enough, frictional slope is steeper than the grey line, so that the system always finds the 5 equilibrium state at the residual level of friction, as line (1). Besides, when Dc is large 6 enough such as case (2), the equilibrium state is found with a small amount of stress 7 drop and fault slip. As a result, stress concentration in the surrounding is large in case 8 (1) comparing to case (2). However the fault character evolves independently from the 9 fault slip and stress drop in Equation (17).

Fault zone  $\Gamma(x,y)$ Z undeformed injection V host rock × x Fault core drained damaged boundary permeable host rock Shear rupturing condition fault core  $\phi(x,y)$  $\kappa(x,y)$ 1 < Central impermeable surrounding rock ultracataclasite  $\Delta u(x,y)$ Fault core width Foliated zone layer h(z,y)modified after Chester et al. (1993)



1 Figure 2 :





1 Figure 4:





























1 Figure 11 :













1 Figure A4 :



1 Figure A5 :

