

Implementation into earthquake sequence simulations of a rate- and state-dependent friction law incorporating pressure solution creep

H. Noda

Department of Mathematical Science and Advanced Technology, Japan Agency for Marine-Earth Science and Technology, Yokohama, Japan.
E-mail: hnoda@jamstec.go.jp

Accepted 2016 February 8. Received 2016 February 1; in original form 2015 November 11

SUMMARY

Pressure solution creep (PSC) is an important elementary process in rock friction at high temperatures where solubilities of rock-forming minerals are significantly large. It significantly changes the frictional resistance and enhances time-dependent strengthening. A recent microphysical model for PSC-involved friction of clay–quartz mixtures, which can explain a transition between dilatant and non-dilatant deformation (d–nd transition), was modified here and implemented in dynamic earthquake sequence simulations. The original model resulted in essentially a kind of rate- and state-dependent friction (RSF) law, but assumed a constant friction coefficient for clay resulting in zero instantaneous rate dependency in the dilatant regime. In this study, an instantaneous rate dependency for the clay friction coefficient was introduced, consistent with experiments, resulting in a friction law suitable for earthquake sequence simulations. In addition, a term for time-dependent strengthening due to PSC was added which makes the friction law logarithmically rate-weakening in the dilatant regime. The width of the zone in which clasts overlap or, equivalently, the interface porosity involved in PSC plays a role as the state variable. Such a concrete physical meaning of the state variable is a great advantage in future modelling studies incorporating other physical processes such as hydraulic effects. Earthquake sequence simulations with different pore pressure distributions demonstrated that excess pore pressure at depth causes deeper rupture propagation with smaller slip per event and a shorter recurrence interval. The simulated ruptures were arrested a few kilometres below the point of pre-seismic peak stress at the d–nd transition and did not propagate spontaneously into the region of pre-seismic non-dilatant deformation. PSC weakens the fault against slow deformation and thus such a region cannot produce a dynamic stress drop. Dynamic rupture propagation further down to brittle–plastic transition, evidenced by geological observations, would require even smaller frictional resistance at coseismic slip rate, suggesting the importance of implementation of dynamic weakening activated at coseismic slip rates for more realistic simulation of earthquake sequences. The present models produced much smaller afterslip at deeper parts of arrested ruptures than those with logarithmic RSF laws because of a more significant rate-strengthening effect due to linearly viscous PSC. Detailed investigation of afterslip would give a clue to understand the deformation mechanism which controls shear resistance of the fault in a region of arrest of earthquake ruptures.

Key words: Numerical solutions; Fracture and flow; Friction; Fault zone rheology; Earthquake dynamics.

1 INTRODUCTION

A fault friction law is one of the key ingredients for the modelling of earthquakes and sequences of them. Laboratory-derived logarithmic rate- and state-dependent friction (RSF) laws (e.g. Dieterich 1979; Ruina 1983) have been widely used in numerical simulations of earthquake sequences (e.g. Rice 1993; Lapusta *et al.* 2000, among many others) in which dynamic rupture propagation and

interseismic stress accumulation accompanied by aseismic fault motion are both simulated in a single framework. In logarithmic RSF laws, shear stress on a fault τ is usually given by

$$\begin{aligned}\tau(V, \theta) &= \sigma_e^{\text{eff}} (\mu_* + a \ln(V/V_*) + b \ln(\theta/\theta_*)), \\ d\theta/dt &= \varphi(V, \theta),\end{aligned}\quad (1)$$

where σ_e^{eff} is the effective normal stress, total normal stress minus pore pressure, V is the slip rate, θ is a state variable, a and b are parameters representing the amount of direct and evolution effects and φ describes the state-evolution law which depends on the version of the RSF law used. For example, among many other variations,

$$\varphi(V, \theta) = 1 - \frac{V\theta}{d_c} \quad (\text{aging law}), \quad (2)$$

and

$$\varphi(V, \theta) = -\frac{V\theta}{d_c} \ln\left(\frac{V\theta}{d_c}\right) \quad (\text{slip law}), \quad (3)$$

where d_c is the state-evolution distance, a characteristic slip scale of evolution of θ (e.g. Ruina 1983). μ_* is the friction coefficient at a reference condition defined by $V = V_*$ and $\theta = \theta_*$ which is usually assumed to be steady-state.

Ruina (1983) defined d_c as the slip scale of exponential decay of linearized perturbations added to steady-state sliding:

$$d_c = -V \left(\frac{\partial \varphi}{\partial \theta} \bigg|_{\text{ss}} \right)^{-1}. \quad (4)$$

d_c in this paper is defined by eq. (4), which is a function of V in general for an arbitrary function φ . Note that d_c in eqs (2) and (3) satisfy eq. (4) regardless of V .

The logarithmic RSF laws are phenomenological laws that can explain many laboratory experiments and a set of the parameters can be determined by fitting the model to experimental data (e.g. Reinen & Weeks 1993; Noda & Shimamoto 2009) for a specific deformation condition. But these parameters have often found to depend on the deformation condition (e.g. Dieterich 1978; Marone *et al.* 1990; Reinen *et al.* 1992; Kawamoto & Shimamoto 1997; Blanpied *et al.* 1998; Saffer & Marone 2003; Niemeijer & Spiers 2005, among many others). Some of those dependencies, as well as the functional form of the RSF laws, were interpreted to give insights into underlying microphysics (e.g. Nakatani 2001; Rice *et al.* 2001; Nakatani & Scholz 2004a,b; Nagata *et al.* 2012) that dictates the frictional resistance and its fluctuation under the experimental condition. Furthermore, some experimental observations were extrapolated to natural conditions and used in modelling natural fault motion. For example, temperature-dependent steady-state rate dependency (e.g. Blanpied *et al.* 1991) was applied to earthquake sequence simulations (e.g. Lapusta *et al.* 2000), and a notion of cut-off velocity of state evolution (e.g. Dieterich 1978) was applied to model slow slip events (e.g. Shibazaki & Shimamoto 2007). Whether the extrapolation is reasonable or not is a tremendously important problem. The dependencies of the parameters on variables which dynamically change during earthquake sequences (e.g. slip rate) urge an update of phenomenological friction laws to microphysics-based ones which are able to express the dependencies, and enable us to upscale the friction laws from laboratory experiments to natural faults.

During an earthquake sequence, the deformation condition such as the slip rate changes spatio-temporally by many orders of magnitude, and the dominant deformation mechanism most likely changes accordingly. For example, among others, the deformation mechanism changes with the strain rate in brittle-plastic transition (e.g. Scholz 1988), and high-velocity coseismic slip causes drastic weakening of a fault by various mechanisms, many of which are triggered by frictional heating (e.g. Sibson 1973; Lachenbruch 1980; Tsutsumi & Shimamoto 1997; Rice 2006; Di Toro *et al.* 2011). Implementation of those mechanisms into numerical simulations of earthquake sequences (e.g. Noda & Lapusta 2010; Shimamoto

& Noda 2014) enables us to discuss earthquake dynamics in terms of fault-materials science and microphysical processes that can be studied in laboratories.

Pressure solution creep (PSC; e.g. Rutter 1976; Shimizu 1995) is one of the important elementary processes in rock friction at a high temperature where solubilities of rock-forming minerals are significantly large. It was experimentally shown that PSC increases the friction coefficient for pure NaCl fault gouge (Bos *et al.* 2000b), decreases the friction coefficient for mixtures of phyllosilicate and NaCl (Bos & Spiers 2002b; Niemeijer & Spiers 2005, 2006, 2007) and enhances time-dependent strengthening of a fault (Bos & Spiers 2002a; Nakatani & Scholz 2004a,b; Niemeijer *et al.* 2008). Microphysical models for the effects of PSC on the frictional behaviour of halite-phyllosilicate and quartz-phyllosilicate mixtures have been developed by Bos & Spiers (2002b), Niemeijer & Spiers (2007) and den Hartog & Spiers (2014). In this study, the recent microphysical model by den Hartog & Spiers (2014, H&S hereafter) is modified to be suitable to be coupled with an elastic medium and implemented into dynamic earthquake sequence simulations using a spectral boundary integral equation method (Lapusta *et al.* 2000).

2 FRICTION LAW ACCOUNTING FOR PSC

2.1 Microphysical model by H&S

In this section, the microphysical model developed by H&S is briefly described. For detailed explanation and the original derivation, please see H&S and den Hartog (2013).

H&S modelled friction experiments on illite-quartz mixtures under hydrothermal conditions around the transition between dilatant and non-dilatant deformation, which is caused by a transition from dilatant slip on anastomosing phyllosilicates operating in competition with compaction due to PSC of quartz clasts, to non-dilatant slip on the phyllosilicates accommodated by shear deformation of the quartz clasts via PSC (Niemeijer & Spiers 2005, 2006, 2007). For convenience, this transition will be denoted by d-nd transition hereafter. Fig. 1 (modified figures from H&S) illustrates the microstructure assumed by H&S for a shear zone consisting of quartz grains and supporting clay matrix. They divided the shear zone into two types of zones, clast body zones and clast overlap zones (width x), and considered pervasive deformation within these zones as well as slip on the wavy foliation anastomosing around the clasts. The slip on the wavy foliation opens up pores at the extensional quartz-matrix interfaces which close by pressure solution compaction creep driven by the effective normal stress σ_n^{eff} .

The shear zone is assumed to have a 2-D body-centred squared lattice structure at zero interface porosity. Note that this assumption is not essential, but reduces one degree of freedom in the parameter space. A rectangular unit cell was considered by H&S which includes 1 quartz grain (Fig. 1a). Given the quartz fraction f_Q and the diameter of the quartz clast grains D , the width of the clast overlap zone at zero interface porosity x_0 and the periodicity parallel to the shear zone L were expressed as (eqs 2 and 1 in H&S)

$$x_0 = D \left(1 - \sqrt{\frac{\pi}{8f_Q}} \right), \quad (5)$$

and

$$L = 2(D - x_0), \quad (6)$$

respectively. Note that f_Q must be in a range $\pi/8 \leq f_Q \leq \pi/4$ in order for the clast overlap zones to exist and for the clasts not to touch

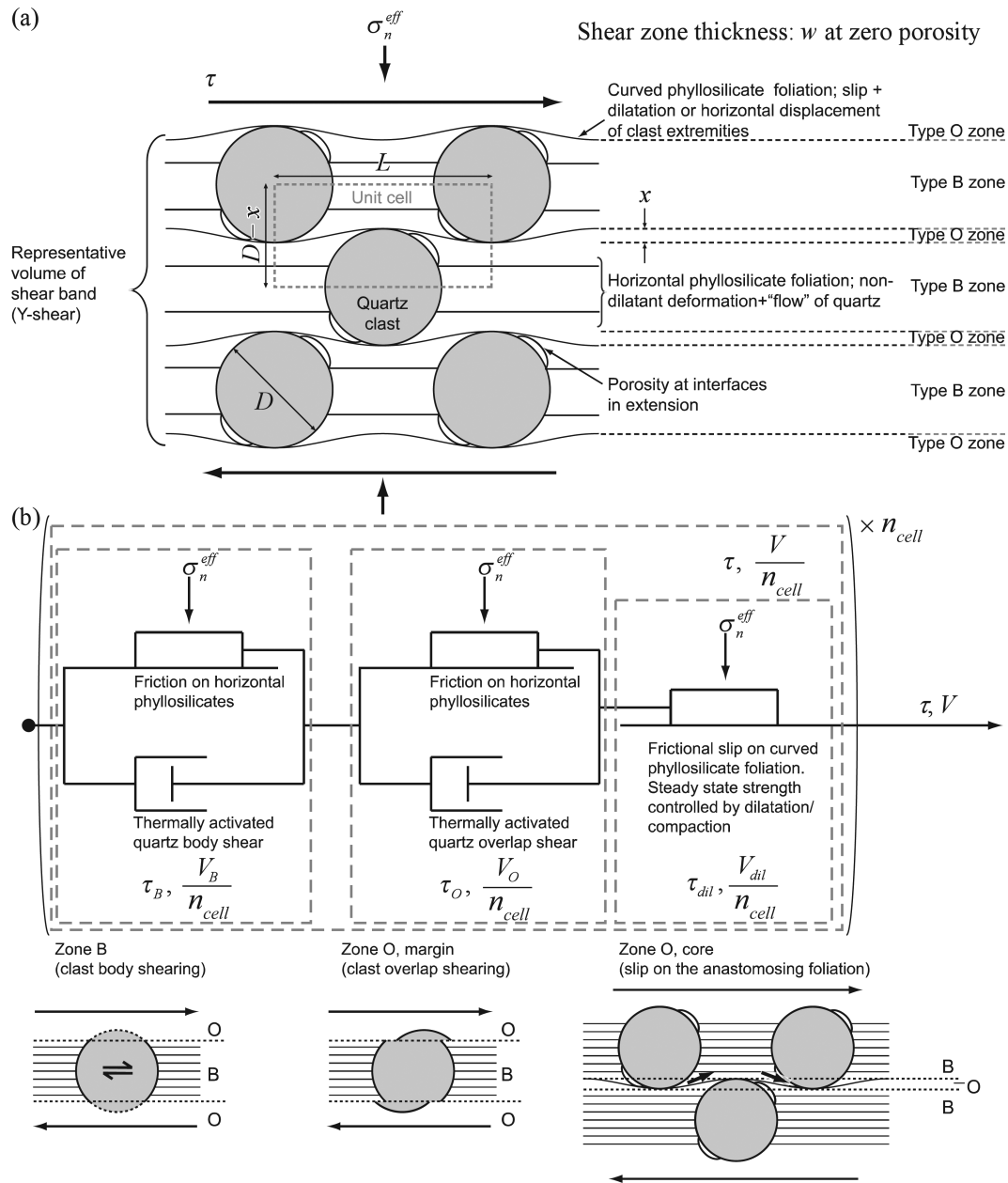


Figure 1. (a) Microstructure assumed by H&S for a shear zone composed of a clay–quartz mixture. Modified from fig. 3(a) in H&S by changing the amplitude of wavy foliations to the true scale. (b) A mechanical analogue diagram indicating an idealization of the clay–quartz mixture illustrated in panel (a). Modified from fig. 4 in H&S.

each other, as mentioned by H&S. The fault-normal dimension of the unit cell is $D - x$ which takes a value of $D - x_0$ at the zero interface porosity limit. w denotes the width of the shear zone at zero interface porosity so that the number of the unit cells across the shear zone n_{cell} is given by

$$n_{cell} = \frac{w}{D - x_0}. \quad (7)$$

H&S assumed that the net slip rate within the shear zone V is partitioned into three types of deformation elements: clast body zones (subscript B), clast overlap zones (subscript O) and dilatant deformation (subscript dil) accommodated by slip on the wavy foliation (Fig. 1b). Note that the former two types of elements account for

linearly viscous PSC of the quartz clasts. The slip rate contributions by the three types of the elements are denoted as V_B , V_O and V_{dil} :

$$V = V_B + V_O + V_{dil}. \quad (8)$$

The elements have different constitutive laws, and support the same applied stresses τ , that is,

$$\tau = \tau_B(V_B, x) = \tau_O(V_O, x) = \tau_{dil}(V_{dil}, x). \quad (9)$$

x is a state variable for which an equation of time evolution was derived by H&S.

The constitutive law for the clast body zone was derived by H&S and takes the form (eq. 5 in H&S):

$$\tau_B(V_B, x) = F_B(x)\mu_B\sigma_n^{eff} + \eta_B(x)V_B, \quad (10)$$

where μ_B is the friction coefficient of clay matrix in the clast body zone, $F_B(x)$ is a geometrical factor and $\eta_B(x)$ is a factor representing viscosity [(stress)/(speed)] of thermally activated PSC. Dissolution/precipitation-controlled PSC of quartz was assumed. Specific expression of these factors and similar ones appearing later in this section were derived by H&S and are shown in Appendix A for completeness.

Similarly, the constitutive law of the clast overlap zone was derived by H&S and takes a form (eq. 6 in H&S):

$$\tau_O(V_O, x) = F_O(x)\mu_O\sigma_n^{\text{eff}} + \eta_O(x)V_O, \quad (11)$$

where μ_O is friction coefficient of clay matrix in the clast overlap zone, $F_O(x)$ is a geometrical factor and $\eta_O(x)$ is a factor representing viscosity [(stress)/(speed)] of thermally activated PSC.

The constitutive law of the dilatant deformation was also derived by H&S (eq. 10 in H&S) and is given by

$$\tau_{\text{dil}}(V_{\text{dil}}, x) = F_{\text{dil}}(x, \mu_{\text{dil}})\mu_{\text{dil}}\sigma_n^{\text{eff}}, \quad (12)$$

where $F_{\text{dil}}(x)$ is a non-dimensional factor modestly larger than unity, accounting for waviness of the foliation. H&S presented a case with rate-independent clay friction. In such a case, μ_{dil} and τ_{dil} do not depend on V_{dil} . This is discussed in the next section where the notation is simplified to $\tau_{\text{dil}}(x)$.

The evolution of x is expressed in terms of V_{dil} and x by H&S as

$$\dot{x}(V_{\text{dil}}, x) = -F_x \frac{x}{D} V_{\text{dil}} + \frac{1}{\eta_{xc}} \left(\frac{D(x_0 - x)}{x_0(D - x)} \right)^n \sigma_n^{\text{eff}}. \quad (13)$$

The first and second terms represent dilatation due to slip on the wavy foliation and time-dependent compaction due to PSC, respectively. F_x is a geometrical factor, η_{xc} is a factor representing viscosity [(stress)/(speed)] of thermally activated pressure solution compaction creep at critical porosity ($x = 0$), and n is a power

exponent that represents the dependency of the compaction rate on the interface porosity. Note that at a steady-state $x = x_{\text{ss}}$ and,

$$\dot{x}(V_{\text{dil}}, x_{\text{ss}}) = 0. \quad (14)$$

At a limit $V_{\text{dil}} = 0$, the first term in eq. (13) is zero, and the second term is positive if $x < x_0$. In this case x would increase with time without slip, which is called aging. However, x cannot exceed x_0 , meaning that this aging and associated strength change saturates at large enough time, since the factors introduced above do not diverge at $x = x_0$.

H&S presented analyses of steady-state behaviour with a constant friction coefficient for μ_B , μ_O and μ_{dil} . In this case, the friction law has zero instantaneous rate dependency during dilatant deformation, as explained in the next subsection. This may cause a problem in simulating the coupling of a fault with the surrounding elastic medium (Rice *et al.* 2001). Therefore, rate-dependent clay friction for μ_B , μ_O and μ_{dil} was used here. The need for a rate-dependent friction for the clay foliation was already recognized by H&S based on the consideration of experimental data. In the following sections, the characteristics of the friction law are presented.

2.2 Instantaneous rate dependency

2.2.1 Rate-independent clay friction

H&S presented analyses of steady-state behaviour with a constant kinetic friction coefficient μ_* for μ_B , μ_O and μ_{dil} . In this case, τ_B and τ_O are linear functions of V_O and V_B , respectively, and τ_{dil} is constant for fixed x . Fig. 2(a) shows those functions as solid lines with the same physical parameters as used by H&S (Table 1) assuming $x = x_0$ as an example. The relation between τ and V is also indicated by the dashed line.

If the applied shear stress τ is smaller than or equal to $\tau_B(0, x)$, all the slip rate contributions are zero, and thus $V = 0$. If

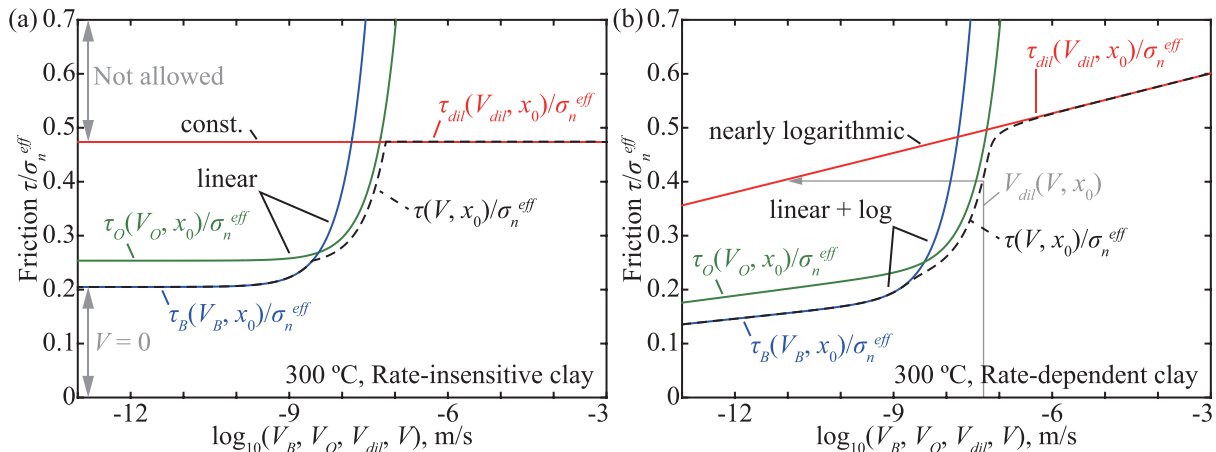


Figure 2. Rate dependency of each element and that of the shear zone at a fixed state variable $x = x_0$. Physical properties are listed in Table 1 unless indicated otherwise. (a) Rate-insensitive clay friction ($a = 0$). (b) Rate-dependent clay friction ($a = 0.01$).

Table 1. Physical parameters used in H&S.

Quartz clast diameter	D	9.3×10^{-7} m
Quartz fraction	f_Q	0.56
Clay friction at reference	μ_*	$0.303 + 5.375 \times 10^{-4} (T - 297.15 \text{ [K]})$
Power exponent of interface porosity	n	0.3
Effective normal stress	σ_n^{eff}	170 MPa
Shear zone thickness	w	5.7×10^{-4} m

$\tau_B(0, x) < \tau \leq \tau_O(0, x)$, then V_B is the only non-zero slip rate contribution. While this regime was included in the mechanical analogue diagram representing the original model by H&S, it was not included in their plots of friction versus V . If $\tau_O(0, x) < \tau < \tau_{\text{dil}}(x)$, then V_B and V_O are non-zero and thus $V = V_B + V_O$.

At $\tau = \tau_{\text{dil}}(x)$, V_B and V_O take their maximum values which are functions of x :

$$\begin{aligned} V_{B\text{max}}(x) &= \frac{\tau_{\text{dil}}(x) - \mu_B \sigma_n^{\text{eff}} F_B(x)}{\eta_B(x)}, \\ V_{O\text{max}}(x) &= \frac{\tau_{\text{dil}}(x) - \mu_O \sigma_n^{\text{eff}} F_O(x)}{\eta_O(x)}. \end{aligned} \quad (15)$$

Kinetic friction determines the maximum shear stress that can be supported by a frictional surface. Thus τ is limited by $\tau_{\text{dil}}(x)$, and larger values are not allowed. The instantaneous stress response to slip rate perturbations at this condition is completely flat as shown in Fig. 2(a).

V_{dil} can be written as a function of V and x by

$$V_{\text{dil}}(V, x) = \max \{0, V - V_{B\text{max}}(x) - V_{O\text{max}}(x)\}. \quad (16)$$

Then the state-evolution law $\varphi(V, x)$ can be constructed by

$$\varphi(V, x) = \dot{x}(V_{\text{dil}}(V, x), x). \quad (17)$$

The relation between τ and V shown in Fig. 2(a) by a dashed line can be constructed for any given value of x between 0 and x_0 . Therefore, this law can be regarded as a RSF law. It should be pointed out that the instantaneously rate-independent characteristic for $V_{\text{dil}} \neq 0$ may cause a problem. H&S demonstrated that the law is rate-weakening in terms of steady-state behaviour in the dilatant regime (i.e. $V_{\text{dil}} \neq 0$). Rice *et al.* (2001) considered interaction of a fault governed by a linearized RSF law and surrounding elastic medium, and analysed the behaviour of linearized perturbation added to a steady-state solution. They showed that if a steady-state rate-weakening fault does not have instantaneous rate dependency, which is the case here, then the perturbation grows irrespective of the wavenumber, and high-wavenumber perturbation propagates near the elastic wave speed. This means that numerical spatial resolution of frictional instability may be quite difficult, and there is no quasi-static regime. As pointed out by Rice *et al.* (2001), both rate- and state-dependencies are necessary in coupling a friction law with the surrounding elastic medium.

2.2.2 Rate-dependent clay friction

H&S (also see den Hartog *et al.* 2012) observed positive steady-state rate dependency in their experiments at temperatures below 250 °C and argued that positive rate dependency of clay friction, which is ubiquitously observed, probably explains such behaviour. Introduction of instantaneous rate-dependent clay friction causes instantaneous rate dependency of the friction law in the dilatant regime. The original model of H&S is therefore modified here to include such rate dependency. In the numerical simulation shown later, similar logarithmic rate dependency on local shear strain rate was assumed for μ_B , μ_O , μ_{dil} . The thickness of the elements are $D - 2x$, x and w_f for the clast body zone, the clast overlap zone and the wavy foliation, respectively. Note that w_f is the local thickness of the foliation measured normal to it, not the amplitude x . In addition, the wavy foliation is oblique to the fault in an absolute sense by $\tan^{-1}(2x/L)$ on average. Then μ_B , μ_O and μ_{dil} are expressed as

$$\mu_B = \mu_* + a \ln \left(\frac{V_B}{n_{\text{cell}}(D - 2x)\dot{\gamma}_*} \right), \quad (18)$$

Table 2. Newly introduced physical parameters for rate-dependent clay friction.

Instantaneous rate dependency of clay	a	0.01
Reference shear strain rate	$\dot{\gamma}_*$	10^{-3} s^{-1}
Thickness of wavy foliation	w_f	10^{-8} m

$$\mu_O = \mu_* + a \ln \left(\frac{V_O}{n_{\text{cell}}x\dot{\gamma}_*} \right), \quad (19)$$

$$\mu_{\text{dil}} = \mu_* + a \ln \left(\frac{V_{\text{dil}}\sqrt{L^2 + 4x^2}}{n_{\text{cell}}Lw_f\dot{\gamma}_*} \right). \quad (20)$$

μ_* is the friction coefficient at a reference shear strain rate $\dot{\gamma}_*$, a represents the logarithmic instantaneous rate dependency of clay.

The newly introduced physical parameters are listed in Table 2. den Hartog *et al.* (2012) reported the instantaneous rate dependency of illite-rich gouge (fig. 3b there). At temperatures below 300 °C, it takes the value around 0.01 with some data scattering and does not show a clear temperature dependency, despite the notion that the instantaneous rate dependency should be proportional to absolute temperature (Nakatani 2001; Rice *et al.* 2001) based on consideration of a microscopic slip process at asperity contacts which requires thermal activation. In the current study, $a = 0.01$ is adopted which is independent of temperature. $\dot{\gamma}_*$ was assumed to be 10^{-3} s^{-1} , which corresponds to typical experimental conditions (a slip rate of $1 \mu\text{m s}^{-1}$ for a 1 mm thick shear zone). $w_f = 10^{-8} \text{ m}$ was used, which is roughly 10 times as thick as the spacing of (001) planes of illite.

Fig. 2(b) is a similar plot to Fig. 2(a), but with rate-dependent clay friction. The relation between τ and V was calculated by numerically solving eq. (8) and (9). They were rearranged and non-dimensionalized to

$$\begin{aligned} 1 - \tau_O(V_O, x)/\tau_B(V_B, x) &= 0, \\ 1 - \tau_O(V_O, x)/\tau_{\text{dil}}(V_{\text{dil}}, x) &= 0, \\ (V_B + V_O + V_{\text{dil}})/V - 1 &= 0. \end{aligned} \quad (21)$$

If V and x are specified, there are three equations with three unknowns (V_B , V_O , V_{dil}). This set of equations was numerically solved with the Newton-Raphson method for a 3-D function. Since the slip rate contributions vary logarithmically, $\log(V_B)$, $\log(V_O)$, $\log(V_{\text{dil}})$ were treated as the variables. This treatment prevents the slip rate contributions from taking negative numbers, making the numerical convergence from an initial guess much more robust.

The increasing functions for the clay friction (eqs 18 to 20) being incorporated, τ becomes an increasing function with respect to V as shown in Fig. 2(b) by the dashed line. V_{dil} can be regarded as a function of V and x similarly to the case with rate-independent clay friction. Its analytic expression is difficult, but graphical expression is quite straightforward (grey arrow in Fig. 2b). The state-evolution law $\varphi(V, x)$ is then defined by eq. (17). Therefore, the friction law by H&S becomes a RSF law with a positive direct effect after introduction of positive rate dependency for clay friction.

2.3 Steady-state behaviour and introduction of a b -term

Fig. 3(a) shows the steady-state friction coefficient with rate-independent clay friction, reproduced from fig. 5(a) in H&S and corrected by adding a regime for $V_O = 0$. The steady-state rate dependency approaches zero at low enough and high enough slip

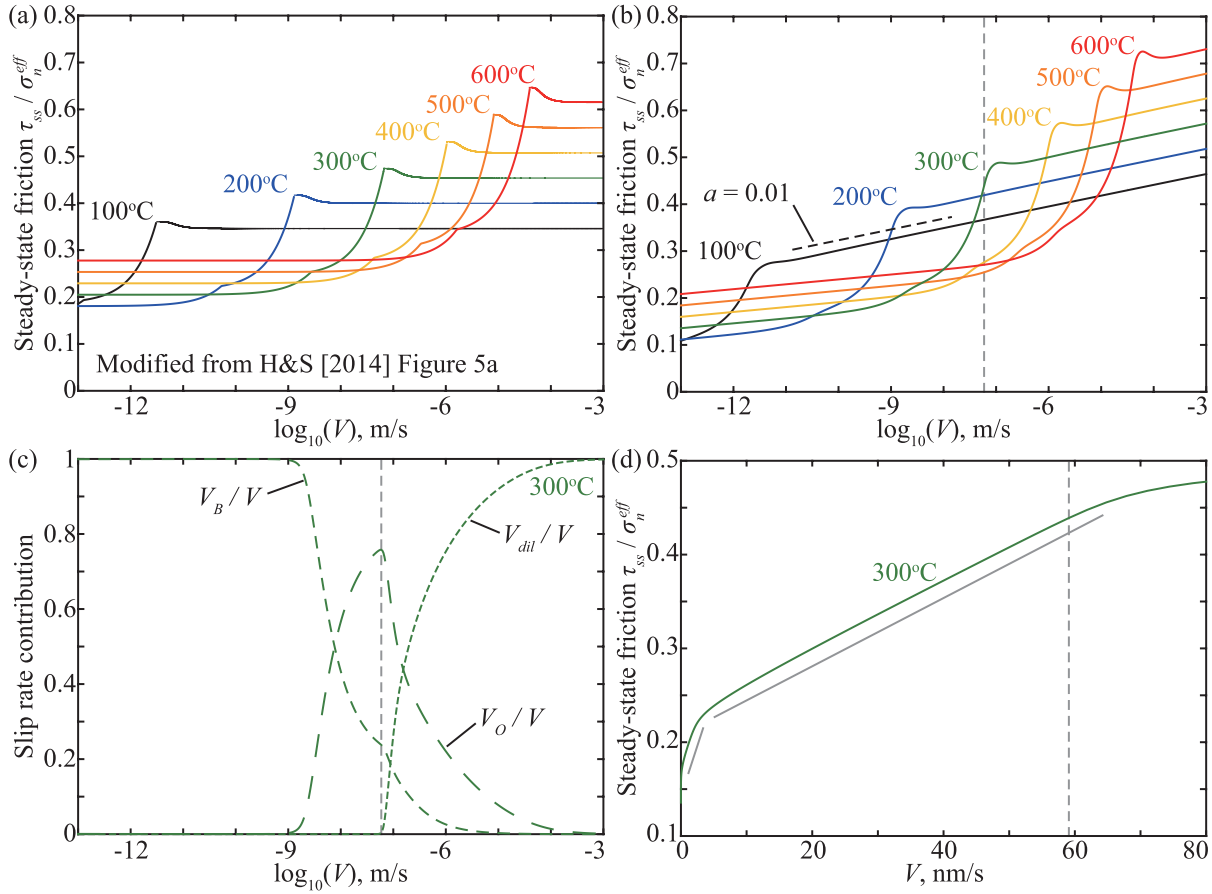


Figure 3. Steady-state shear stress and slip rate partitioning for the physical parameters listed in Table 1 at temperatures from 100 °C to 600 °C. (a) Modified figure from fig. 5(a) in H&S [2014] in which the friction coefficient of clay is rate-insensitive ($a = 0$). (b) Steady-state friction coefficients with rate-dependent clay friction. The direct-effect parameter a is 0.01. (c) Slip rate partitioning for 300 °C. (d) A plot for the case with 300 °C in panel (b) with a linear horizontal scale. Two linear parts are recognized (also see Fig. 2b).

rates. At low enough slip rates, the flow stresses of quartz grains are so small that the shear stress is mostly supported by the clay matrix. At the other extreme, dilatant deformation occurs at critical state porosity and the strength is governed by frictional sliding on horizontally aligned clay foliation. The temperature-strengthening tendencies in these regimes are due to the positive dependency of clay friction on T in the experiments cited by H&S (den Hartog & Spiers 2013; see Table 1).

At a steady-state, eq. (14) is satisfied in addition to eqs (8) and (9). Those equations were rearranged and non-dimensionalized to

$$\begin{aligned} 1 - \tau_O(V_O, x_{ss})/\tau_B(V_B, x_{ss}) &= 0, \\ 1 - \tau_O(V_O, x_{ss})/\tau_{dil}(V_{dil}, x_{ss}) &= 0, \\ (V_B + V_O + V_{dil})/V - 1 &= 0, \\ \frac{\dot{x}(V_{dil}, x_{ss})}{-F_x x_{ss} V_{dil}/D} &= 0. \end{aligned} \quad (22)$$

The denominator in the last equation represent dilatancy due to slip on the wavy foliation. If V is specified, there are four equations with four unknowns (V_O , V_B , V_{dil} , x_{ss}). Note that $\tau_{ss} = \tau_B = \tau_O = \tau_{dil}$ at the solution. The set of equations was solved with the Newton-Raphson method for a 4-D function. Similarly to the calculation of Fig. 2(b), $\log(V_B)$, $\log(V_O)$, $\log(V_{dil})$ were treated as variables. If a wide range of V is investigated, x_{ss} becomes very close to 0 at large V , and so does $x_0 - x_{ss}$ at small V . If one adopts $\log(x)$ as the forth variable (logarithmic transformation to avoid negative x), then the

solution search fails due to loss of significant digits at small V . In order to avoid this problem, in every step of the solution search, the smaller of $\log(x_{ss})$ and $\log(x_0 - x_{ss})$ was used as the forth variable to be updated.

Introduction of rate dependency of the clay causes an overall steady-state rate-strengthening trend (Fig. 3b). At a large V limit, the rate dependency $d\mu_{ss}/d\ln(V)$ is given by a , which agrees better with the experimental results by H&S than their original analyses with rate-independent clay friction. In addition, gradual onset of the dilatant deformation makes the d-nd transition smooth which would benefit numerical simulations.

The slip rate contributions from the 3 types of elements are shown in Fig. 3(c) for the steady-state condition at 300 °C. The clast body zone has the largest slip rate contribution at small V . With increasing slip rate, the contribution of the clast overlap zone V_O increases and becomes the largest at 8.76 nm s⁻¹. This transition causes a change in the viscosity as shown in Fig. 3(d). Because of the smooth onset of shear deformation of the clast overlap zone, this change is gradual, while rate-independent clay friction causes a singular point (see Fig. 2).

V_O/V reaches its maximum value at 59.1 nm s⁻¹ (vertical grey dashed line in Figs 3b–d), and then decreases. At this point, the contribution of the dilatant deformation V_{dil} starts to become significant and $\tau_{ss}(V)$ deviates from a linear trend (Fig. 3d). Note that rate-independent friction caused a singular on-off behaviour of the dilatant deformation. This makes the deviation from the

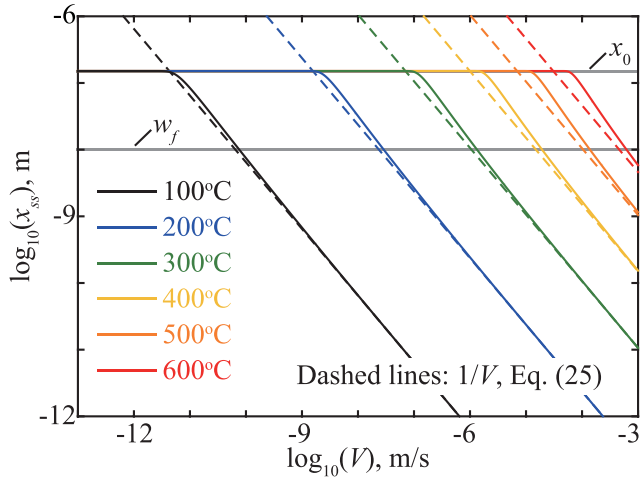


Figure 4. The state variable at steady-state x_{ss} with the physical parameters listed in Table 1 at different temperatures. Dashed lines represent $1/V$ dependency given by eq. (25). Grey horizontal lines represent the maximum value x_0 and the thickness of wavy foliation w_f .

linear trend singular even without state-evolution (Fig. 2a). With a rate-dependent clay friction, the friction law consists of smooth functions, and this deviation is also smooth.

As shown here, the constitutive behaviour is quite complex. The model by H&S was constructed in order to explain the positive-negative-positive trend in the steady-state rate dependency around the d-nd transition. However, there is no experimental evidence for the transition associated with the change in viscosity at the onset of deformation within clast overlap zone (Fig. 3d) to the author's knowledge. It may be possible to reduce the number of elements in the microphysical model for simplicity while keeping the ability to express the d-nd transition, but I would like to stick to the published model by H&S.

Fig. 4 represents the steady-state state variable x_{ss} . It stays close to its maximum value x_0 , and then decreases approximately as $1/V$ as V increases. If $x \ll x_0$, eq. (13) can be approximated by

$$\dot{x}(V_{\text{dil}}, x) \approx -F_x \frac{x}{D} V_{\text{dil}} + \frac{\sigma_n^{\text{eff}}}{\eta_{xc}}. \quad (23)$$

At steady state (eq. 14),

$$x_{ss} \approx D \frac{\sigma_n^{\text{eff}}}{\eta_{xc} F_x V_{\text{dil}}(V, x_{ss})}. \quad (24)$$

In addition, if $V \approx V_{\text{dil}}$, which is the case at high enough slip rate as shown in Fig. 3(c), then

$$x_{ss} \approx D \frac{\sigma_n^{\text{eff}}}{\eta_{xc} F_x V}. \quad (25)$$

Note that x_{ss} approaches zero as V increases (critical state behaviour). The dashed lines Fig. 4 are estimates given by eq. (25) which give an excellent fit to the calculated x_{ss} at high enough slip rates.

It should be mentioned that as V increases the value of x becomes smaller than the assumed thickness of the shear zone on wavy foliation w_f or even smaller than the basal (001) spacing of illite. Since the wavy foliation is assumed to be within the clast overlap zone (Fig. 1), such a small value of x is unrealistic. The model by H&S successfully explained the d-nd transition and associated change in the sign of rate dependency qualitatively. But its

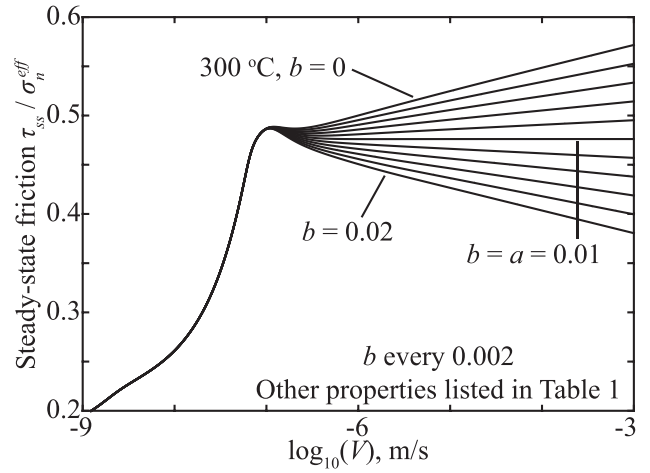


Figure 5. Steady-state friction for different values of b ranging from 0 to 0.02 at 300 °C. The other physical parameters are listed in Table 1.

direct applicability to deformation well inside the dilatant regime was not considered, where conventional logarithmic RSF laws (e.g. Dieterich 1979; Ruina 1983) were established. For example, the steady-state rate dependency in the dilatant regime in Fig. 3(b) is eventually given by the assumed a for the clay friction. This indicates the absence of an evolution effect in the dilatant regime, which contradicts with experimental results by den Hartog *et al.* (2012) for illite-rich gouge.

Pure phyllosilicate has evolution effect [e.g. pure muscovite by den Hartog *et al.* (2013), and near-pure (94 per cent) illite by den Hartog *et al.* (2014)], and Y-shear foliation ($x \ll x_0$) may be approximated by pure illite. The experiments of the near-pure illite yielded the steady-state rate dependency from -0.005 to 0 at 350 °C. The enhancement of rate-weakening by the onset of dilatant deformation is significant in a rather narrow range of deformation conditions and cannot solely explain the experimentally observed wider range of deformation conditions for rate-weakening (see figs 1 and 7 in H&S). In the earthquake sequence simulations shown later, the clay friction of the wavy foliation, which dictates the shear resistance well inside the dilatant regime, was modified by adding a term for the state-evolution effect in the logarithmic RSF laws,

$$\mu_{\text{dil}} = \mu_* + a \ln \left(\frac{V_{\text{dil}} \sqrt{L^2 + 4x^2}}{n_{\text{cell}} L w_f \dot{\gamma}_*} \right) + b \ln \left(\frac{x}{x_0} \right), \quad (26)$$

motivated by the approximate $1/V$ dependency of x_{ss} on V (eq. 25). It has to be mentioned that the assumed constitutive law for clay, particularly the b -term, is phenomenological although the effect of PSC was introduced with a concrete physical basis.

Fig. 5 shows steady-state friction as a function of V at 300 °C for different values of b from 0 to 0.02. The case with $b = 0$ is identical to one of the cases plotted in Fig. 3(b). The high-velocity part is approximately rate-neutral with $b = a = 0.01$ and shows logarithmic rate-weakening with $b > a$, being similar to logarithmic RSF laws (e.g. Ruina 1983).

2.4 State-evolution slip distance d_c

In the present friction law, the characteristic state-evolution slip distance d_c defined by linearization of the friction law (e.g. eq. 4) is not an externally given parameter like for conventional RSF laws, but a function of the assumed microphysical parameters and deformation

conditions. It may change with time as slip rate changes during earthquake sequences. In simulating them, an adaptive control on the length of a time step is inevitable (e.g. Lapusta *et al.* 2000), and estimation of d_c and a corresponding timescale of state-evolution is required at every computational time step. We can define the rate-dependencies (A and $A - B$) by

$$A = V \left. \frac{\partial \tau}{\partial V} \right|_{ss}, \quad (27)$$

and

$$A - B = V \frac{d\tau_{ss}}{dV}, \quad (28)$$

and d_c by,

$$d_c = -V \left(\left. \frac{\partial \varphi}{\partial x} \right|_{ss} \right)^{-1}, \quad (29)$$

similarly to eq. (4). They can be used to determine the maximum time step allowed to avoid numerical instability at least for near-steady-state situations (Lapusta *et al.* 2000).

The black lines in Figs 6(a) and (b) show d_c and the corresponding timescale d_c/V , respectively, as a function of V , using the physical parameters listed in Tables 1 and 2, with $b = 0.01$ ($=a$), and at $T = 300^\circ\text{C}$. In the non-dilatant regime, d_c is smaller for a smaller value of V , and so is d_c/V . Note that d_c/V vanishes as V approaches zero. This means that a shorter time step is required for temporal resolution of state-evolution during slower deformation. In earthquake sequence simulations, long interseismic periods must be simulated with much longer time steps than short coseismic periods because of restricted computational resources. Therefore, the vanishing timescale d_c/V is a critical problem.

The derivative of φ with respect to x is

$$\frac{\partial \varphi}{\partial x} = \frac{\partial \dot{x}}{\partial x} + \frac{\partial \dot{x}}{\partial V_{dil}} \frac{\partial V_{dil}}{\partial x}, \quad (30)$$

and from eq. (13), the first term is given by

$$\frac{\partial \dot{x}}{\partial x} = -F_x \frac{1}{D} V_{dil} - \frac{1}{\eta_{xc}} \frac{n D^n (D - x_0) (x_0 - x)^{n-1}}{x_0^n (D - x)^{n+1}} \sigma_n^{\text{eff}}. \quad (31)$$

This diverges for $n < 1$ at the limit where $x_0 - x$ approaches zero, which causes the problem of the vanishing timescale. H&S adopted $n = 0.3$. Figs 6(a) and (b) represent the effect of n and show that a larger n causes a milder decrease in d_c , and $n = 1$ yields almost constant d_c/V at low enough slip rates (an inset in Fig. 6b), which makes simulations of interseismic processes much easier. Note that for conventional RSF laws with a constant d_c , the timescale would increase as V decreases proportionally to $1/V$, allowing longer time steps for slower slip processes.

$n = 0.3 < 1$ was selected by H&S in order to express a large length-to-width ratio of the pores at the onset of dilatation (i.e. at the debonding of the interface) and a decrease in it as the pores expand. But the resulting rapid compaction at low interface porosity cannot be resolved in earthquake sequence simulations. This problem could be resolved by introducing, for example, truncation of state-evolution at low porosity or a minimum size of pores comparable to the thickness of an absorbed water layer. This point deserves further study. In the simulations shown later, $n = 1$ is adopted.

$\tau_{ss}(V)$ with different n from 0.3 to 2 overall traces each other (Fig. 6c). At the d-nd transition, a larger n causes a less significant peak (Fig. 6c). It should be pointed out that the rate-independent clay friction used by H&S predicts a jump in the rate dependency while

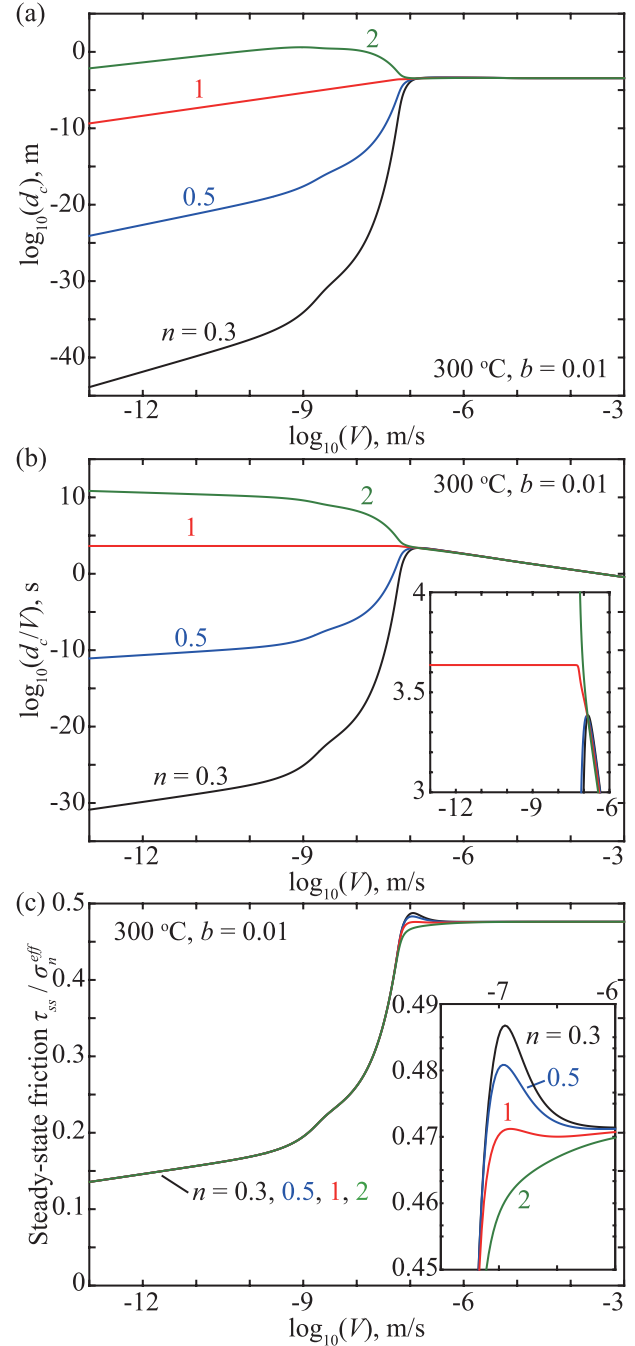


Figure 6. (a) d_c derived from linearization of the friction law around steady state (eq. 29) at $T = 300^\circ\text{C}$, with $b = 0.01$, and for various n . The other parameters are listed in Table 1. (b) A timescale of state evolution d_c/V for the same conditions as in panel (a). For $n < 1$, the timescale vanishes as V approaches zero. An inset shows that the timescale stays almost constant with V at $n = 1$. (c) The effect of n on the steady-state friction coefficient. An inset is a magnified view around the d-nd transition.

the experimental results show a rather smooth change in the rate dependency fig. 1a in H&S. The choice of $n = 1$ reduces the peak in the plot of $\tau_{ss}(V)$ in better agreement with the experiments discussed by H&S but also produces less-marked velocity weakening than seen in the same experiments.

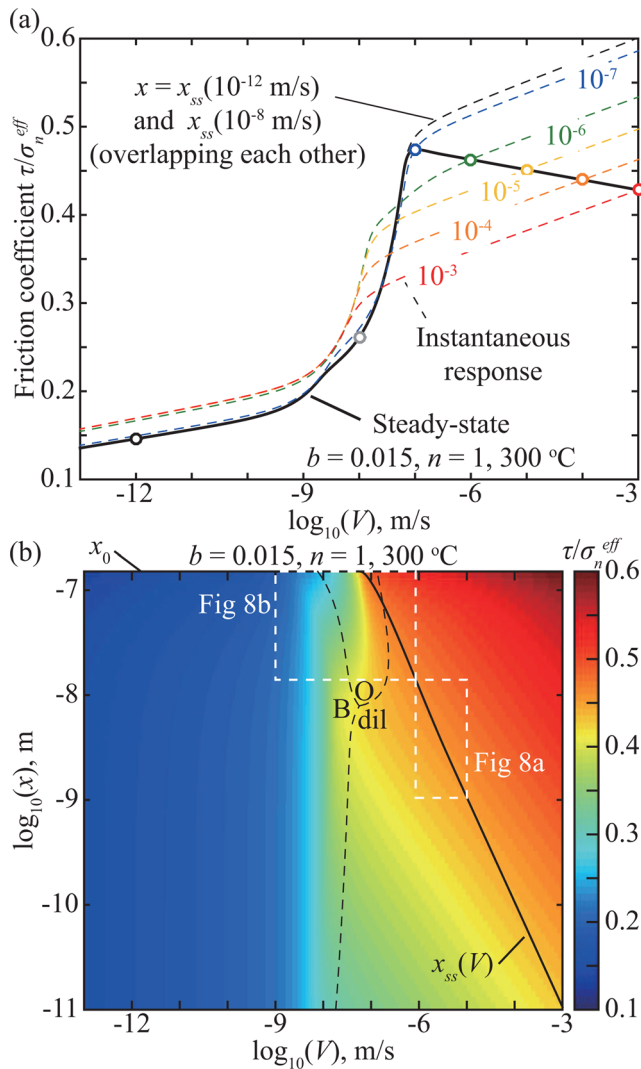


Figure 7. (a) Steady-state (the thick black line) and instantaneous (dashed coloured lines) responses of the friction coefficient with $b = 0.015$ ($a - b = -0.005$) and at 300°C . x of each instantaneous dependency is the steady-state value at the labelled slip rate indicated by circles of the same colour. (b) A colour plot of the friction coefficient as a function of V and x . The black line indicates steady state. Black dashed lines divide the plot area into 'O', 'B' and 'dil' regimes defined by the maximum of the slip rate contributions V_O , V_B and V_{dil} , respectively. White dashed squares are trajectories of simulated velocity steps in Figs 8(a) and (b).

2.5 Transient behaviour

Fig. 7(a) shows the steady-state friction coefficient as a function of V as a thick black line, and instantaneous responses in dashed coloured lines which were calculated by solving eq. (21). $b = 0.015$ and $n = 1$, and $T = 300^\circ\text{C}$ were assumed and the other parameters are listed in Tables 1 and 2. In the dilatant regime, a series of straight lines for the instantaneous responses with positive slope intersect the steady-state line. This is exactly similar to the characteristic of conventional logarithmic RSF laws (e.g. fig. 3 in Ruina 1983). Fig. 7(b) shows the friction coefficient $\tau/\sigma_n^{\text{eff}}$ in a (V, x) space. The space was divided by black dashed lines delineating the largest slip rate contributions, and x_{ss} as a function of V is indicated by a black solid line. A velocity step is expressed by a trajectory parallel to the

coloured dashed lines in Fig. 7(a) and a horizontal line in Fig. 7(b), and state-evolution at a constant slip rate is expressed as a vertical line in both figures.

Mechanical responses following ideal velocity steps are simulated to see the transient behaviour of the present friction law. Initially, a steady-state at a certain slip rate was assumed, and then V was abruptly changed. After a velocity step, the state variable was evolved by integrating eq. (13) in time with a second-order accurate exponential time-differencing method with a constant time step $0.01 d_c/V$ where d_c was estimated at the post-step slip rate.

Velocity-steps in the dilatant regime are characterized by peak-decay behaviour (Fig. 8a) similarly to conventional logarithmic RSF laws. In this regime, large x and thus small interface porosity correlates with large τ .

The opposite is the case when the clast body zones control has the largest slip rate contribution ($V < 10^{-8}$ in Fig. 7); large x causes small τ . In the microphysical model adopted here, a thinner clast overlap zone is associated with a higher porosity, causing a larger shear resistance. In the author's opinion, the correlation between porosity and shear resistance is a bit counter-intuitive, and experimental validation of this point and a potential update of the microphysical model may deserve future study. Note that as long as deformation is applied in the non-dilatant regime, the instantaneous responses and the steady-state line trace almost each other (Fig. 7a), and the friction law may be approximated well by a pure rate-strengthening law. In the non-dilatant regime, significant porosity cannot be generated because x_{ss} is very close to x_0 . The porosity effect discussed here is the case when a fault experiences transition from the dilatant regime to the non-dilatant regime, but the porosity collapses rapidly since d_c (Fig. 6a) is vanishingly small in the non-dilatant regime.

Because of the different correlations between τ and x in the dilatant and non-dilatant regimes, velocity-steps across the d-nd transition show qualitatively non-symmetric behaviour (Fig. 8b). The upward step into the dilatant regime yields a peak-decay behaviour similar to conventional RSF laws, while the downward step into the non-dilatant regime yields a monotonic decay to the steady-state value with a much shorter state evolution distance d_c (see the last section). Velocity steps down to the transitional regime (V from $10^{-8.5}$ m s $^{-1}$ to 10^{-7} m s $^{-1}$) are shown in Figs 8(c) and (d). They indicate that between the dilatant and non-dilatant regimes, downward velocity-steps show the behaviour which could be 'explained' by a logarithmic RSF law with 2 state variables with positive and negative b -values, in spite of the fact that the present friction law has only one state variable. Such behaviour was experimentally observed at hydrothermal conditions (e.g. Blanpied *et al.* 1998). Indeed, den Hartog *et al.* (2012) sometimes used a negative second state variables to fit a conventional RSF law to their experiments.

In this section, consequences of the model by H&S was analysed and modifications made were explained in order to use the model in earthquake sequence simulations. But it should be pointed out that some of the characteristics shown do not have concrete experimental evidence (e.g. the scaling between d_c and V , and complex behaviour after velocity steps). Comparisons of experiments and microphysical models not only in steady-state characteristics, but also in transient behaviours are important future study towards improvement in the microphysical models and physics-based fault constitutive laws.

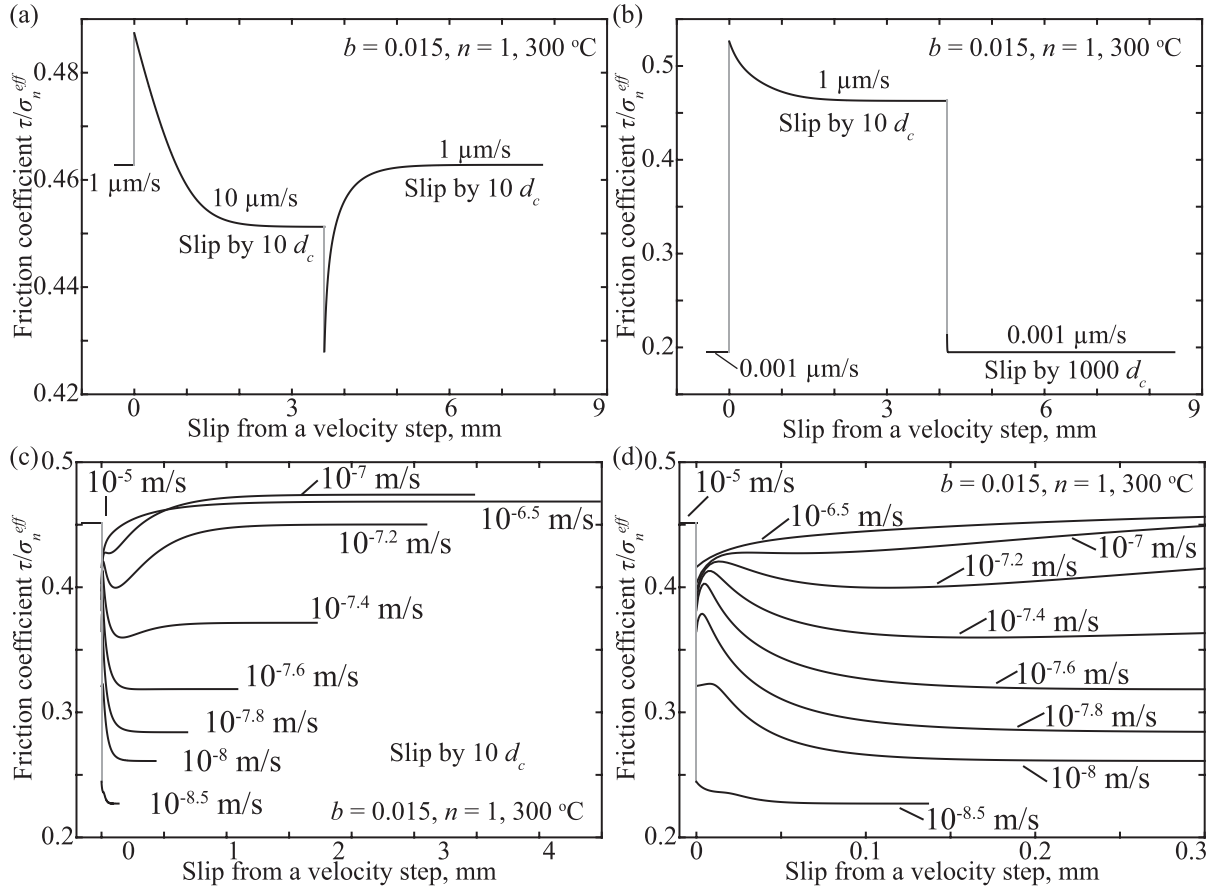


Figure 8. Simulated velocity steps with $b = 0.015$ ($a - b = -0.005$) and at 300°C . The other parameters are listed in Table 1. The grey lines represent abrupt jumps. The amount of simulated slip after the steps are labelled. (a) Velocity steps within the dil-regime, between $1\ \mu\text{m s}^{-1}$ and $10\ \mu\text{m s}^{-1}$. (b) Interregime velocity steps, showing qualitatively non-symmetric behaviour. (c) Downward velocity steps from the dil-regime to O- or B-regimes. (d) A magnified view of panel (c) closely after the velocity steps.

3 METHODOLOGY OF EARTHQUAKE SEQUENCE SIMULATIONS

3.1 Solution techniques

The friction law explained and investigated in Section 2 was implemented in 2-D earthquake sequence simulations using a spectral boundary integral equation method (Lapusta *et al.* 2000). In this method, shear stress on a fault τ is expressed as

$$\tau = \tau_0 + \phi - \frac{G}{2c_s} V = \tau_0 + \phi - \frac{G}{2c_s} (V_B + V_O + V_{\text{dil}}), \quad (32)$$

where τ_0 is the shear stress on the fault which would be achieved if there were no slip on the fault, ϕ is a functional term depending on the previous spatio-temporal distribution of the slip rate V and on the current distribution of the slip, c_s is the shear wave speed, and G is the shear modulus.

The solution technique is overall identical to Noda & Lapusta (2010); the functional term is calculated with a mid-point rule, the state variable x (or $x_0 - x$ if $x > x_0/2$) is updated with a second-order accurate exponential time-differencing method, and the time step is varied adaptively following Lapusta *et al.* (2000) with a safety factor of 2. There are a few differences in detail which may be worth noting.

At every time step x and ϕ are updated, and then eqs (9) and (32) were solved simultaneously for V_B , V_O and V_{dil} by looking for a

solution to the following non-dimensionalized equations:

$$\begin{aligned} 1 - \left(\tau_0 + \phi - \frac{\mu}{2c_s} (V_B + V_O + V_{\text{dil}}) \right) / \tau_B(V_B, x) &= 0, \\ 1 - \left(\tau_0 + \phi - \frac{\mu}{2c_s} (V_B + V_O + V_{\text{dil}}) \right) / \tau_O(V_O, x) &= 0, \\ 1 - \left(\tau_0 + \phi - \frac{\mu}{2c_s} (V_B + V_O + V_{\text{dil}}) \right) / \tau_{\text{dil}}(V_{\text{dil}}, x) &= 0. \end{aligned} \quad (33)$$

Again, the Newton-Raphson method for a 3-D function was used, while treating $\log(V_B)$, $\log(V_O)$ and $\log(V_{\text{dil}})$ as the variables to be updated. Once the solution search converged, $(\tau_B + \tau_O + \tau_{\text{dil}})/3$ was adopted as the estimate of τ . Even if an explicit analytical expression for $\tau(V, x)$ is unknown, it is possible to implement an implicit form of the friction law into earthquake sequence simulations.

In the adaptive stepper proposed by (Lapusta *et al.* 2000), the rate-dependencies (A and $A - B$) and the state-evolution distance d_c (eqs 27–29) are used. Those values are constants for many conventional RSF laws, but vary in the present friction law as demonstrated in previous sections. They are estimated every time step at all nodes around the steady-state at the current value of V .

3.2 Model setting

Similarly to Lapusta *et al.* (2000), a strike-slip fault was assumed and discretized in depth (z -direction). This is an anti-plane problem

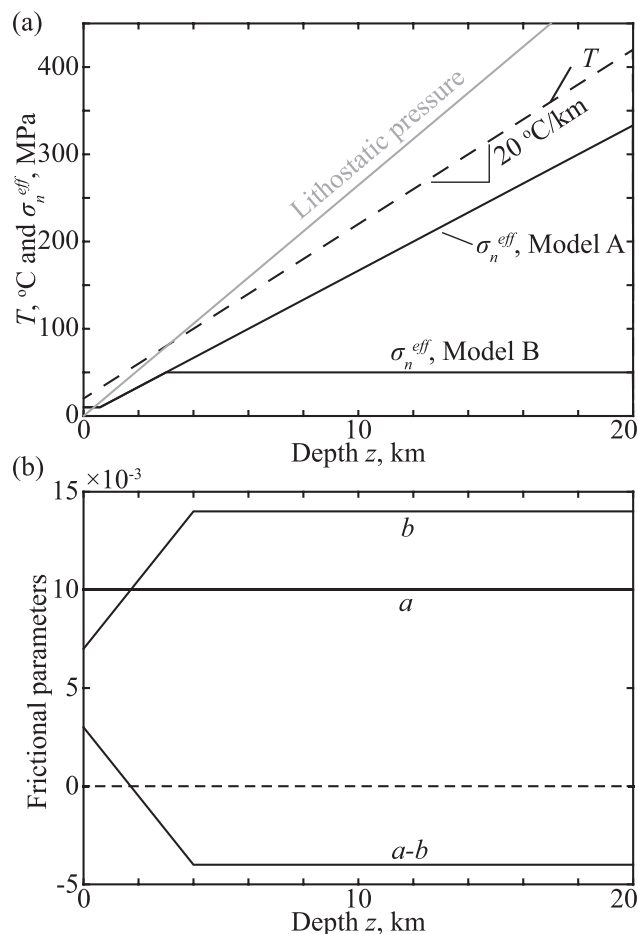


Figure 9. Model setting of the present earthquake sequence simulations. (a) Distributions of T (dashed line) and σ_n^{eff} (solid lines). Lithostatic pressure distribution is also shown by a grey line, indicating the assumed pore pressure distribution at depth. (b) Parameters a , b and $a - b$.

with a mirror located at the ground surface. A periodic boundary condition, which is required in the adopted method, is assumed by setting another mirror at the depth of $z = 25.6$ km. The deep portion of the fault $z < 20$ km is assumed to be sliding steadily by $V = V_{pl} = 10^{-9}$ m s $^{-1}$.

The host rock had a shear wave speed $c_s = 3$ km s $^{-1}$ and a shear modulus $G = 30$ GPa. The temperature T was assumed to be 20 °C at the ground surface, and to increase with depth by 20 °C km $^{-1}$ (the dashed line in Fig. 9a). Two different distributions of the effective normal stress σ_n^{eff} were tested (solid lines in Fig. 9a). In Model A, a hydrostatic pore pressure distribution and lithostatic total normal stress distribution were assumed so that σ_n^{eff} increases with depth by 16.66 MPa km $^{-1}$. In Model B, excess pore pressure at depth was considered and an upper limit for σ_n^{eff} was set at 50 MPa, similarly to Lapusta *et al.* (2000). In addition, a lower limit for σ_n^{eff} was set at 10 MPa to avoid zero-normal stress at $z = 0$.

It should be noted that complex structure of natural fault zones (e.g. distribution of different fault rocks, grain size distribution, mineral assemblage and strain localization) causes difficulty in estimating relevant values for the physical parameters. They may well be distributed heterogeneously along a fault, and fluctuate with time if the actively deforming shear zone thickness changes during a sequence of earthquakes. Obtaining reasonable constraints on the parameters is an important future study.

Table 3. Physical parameters used in earthquake sequence simulations. Also see Fig. 9. For parameters not listed here, the values in Tables 1 and 2 were used.

Evolution effect in dilatant regime	b	0.007 to 0.014
Shear wave speed	c_s	3 km s $^{-1}$
Quartz clast diameter	D	10 $^{-4}$ m
Quartz fraction	f_Q	0.6
Shear modulus	G	30 GPa
Power exponent of interface porosity	n	1
Shear zone thickness	w	10 $^{-2}$ m

Physical parameters used in the simulations that are different from the previous section are listed in Table 3. In this study, in order to minimize the complexity and to investigate the consequence of the present friction law, most of the parameters were assumed uniform. Some of them were selected in order for the simulations to produce earthquakes in a realistic depth range. The shear zone was assumed to consist of mixture of 60 per cent quartz (diameter $D = 100$ μ m) and 40 per cent clay forming the matrix. The grain size was selected to make the earthquake sequence simulations comparable with a standard model by Lapusta *et al.* (2000) with a logarithmic RSF law so that the effect of implementation of PSC is illuminated. The shear zone width w was assumed to be 1 cm which corresponds to the typical width of an ultra-cataclasite (or clayey gouge if unconsolidated) layer (e.g. Chester & Chester 1998). The rate dependency parameter of clay a was assumed to be 0.01, as in the previous sections, and the amount of the evolution effect inside the dilatant regime b was set 0.007 at the ground surface, increasing linearly with depth, and kept constant at 0.014 below $z = 4$ km (Fig. 9b). In the seismogenic depth range, the rate dependency in the dilatant regime $a - b$ was -0.004 , identical to that in the model by Lapusta *et al.* (2000). Based on the discussion in the previous section, $n = 1$ was assumed.

It should be noted that the stable clay mineral changes with depth [e.g. smectite-illite transition (Elliott & Matisoff 1996) and illite-muscovite transition (Hunziker 1986)]. Implementation of a realistic clay mineralogy distribution and associated clay friction deserves future study. In this study, the constant value $a = 0.01$ was adopted, motivated by the illite data, and the focus was put on the effect of PSC on the fault behaviour.

Fig. 10 shows profiles of fault properties defined around steady-state sliding at $V = V_{pl}$. They are, strictly speaking, not realized in the actual history of fault motion unless the fault creeps at a constant slip rate V_{pl} , but they may be useful in illustrating the basic characteristics of the models.

The strength profiles (i.e. $\tau_{ss}(V_{pl})$) are shown in Fig. 10(a). As discussed in previous studies, there was a peak in the strength near the d-nd transition (e.g. Bos & Spiers 2002b). It was located at the depth of 10.6 km and 12.0 km for Model A and B, respectively, being deeper with a smaller σ_n^{eff} (larger pore pressure). Accordingly, the deeper limit of steady-state rate-weakening ($d\tau_{ss}/d\ln(V) < 0$) at $V = V_{pl}$ was deeper with a smaller σ_n^{eff} (Fig. 10b). Immediately adjacent to the rate-weakening portion, at greater depth, there was a gigantic positive peak in the rate dependency due to the large rate dependency of PSC, which is almost linearly viscous. In addition, there was another peak in the rate-dependencies at the transition corresponding to the change in viscosity (Fig. 3d).

In many simulations using RSF laws, the state-evolution slip distance d_c was the most poorly constrained parameters and thus a parameter study on it was often conducted (e.g. Lapusta & Rice 2003). In the present model, d_c was determined as a consequence of the microphysical model (Fig. 10c). With the parameter values

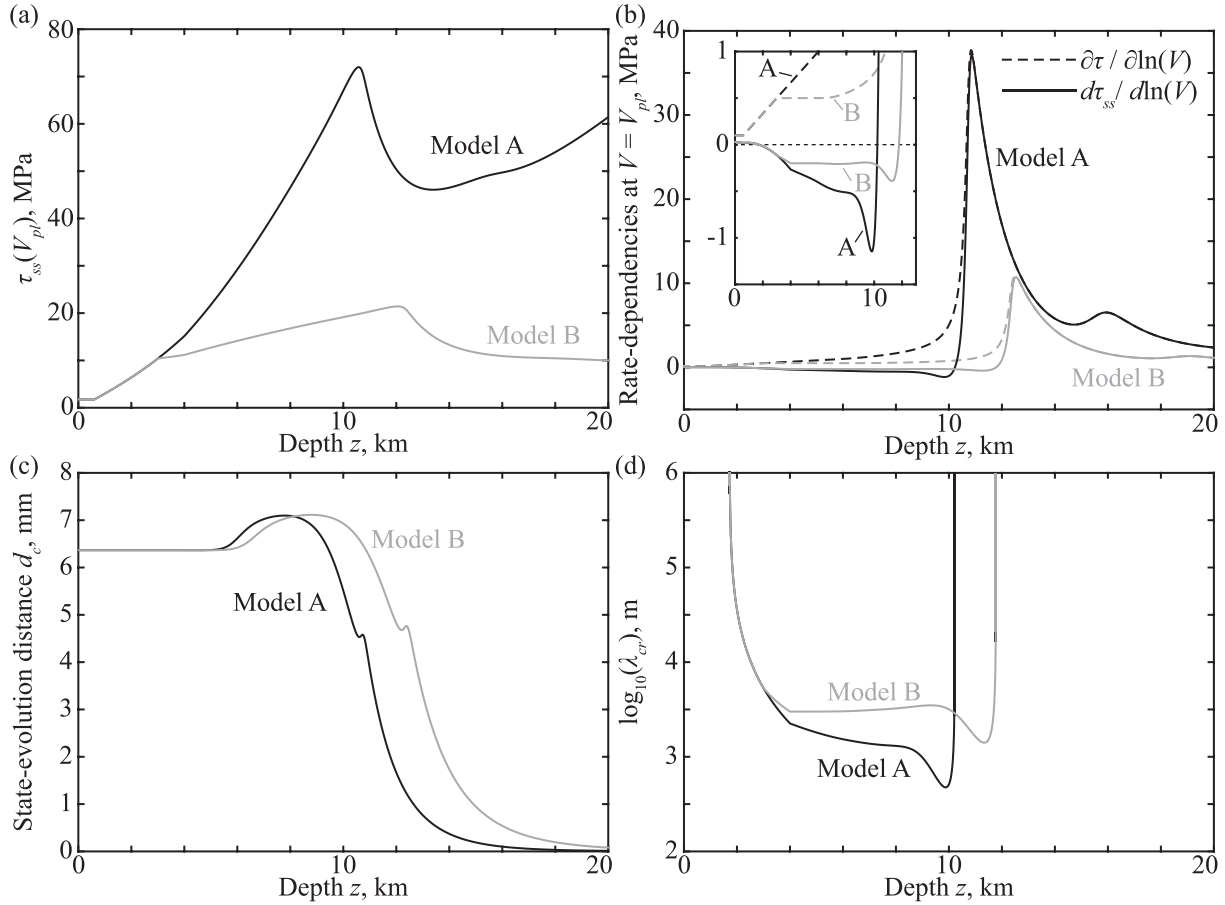


Figure 10. Steady-state and near-steady-state characteristics of the simulated strike slip faults. (a) Strength profile $\tau_{ss}(V_{pl})$, (b) rate dependencies (steady-state dependency in solid and the direct effect in dashed lines), (c) state-evolution slip distance d_c and (d) critical wavelength λ_{cr} . Note that Models A and B have different distribution of σ_n^{eff} (Fig. 9a). The inset in panel (b) is a magnified plot of the rate dependencies in the shallow part of the fault.

adopted here, the critical wavelength λ_{cr} for the impossibility of steady-state sliding at V_{pl} (Rice *et al.* 2001):

$$\lambda_{cr} = -\pi \frac{\mu}{d\tau_{ss}/d\ln(V)} \bigg|_{V_{pl}} d_c, \quad (34)$$

took the minimum value 475 m and 1.41 km for Model A and B, respectively (inset in Fig. 10c).

τ_0 was assumed to be consistent with steady-state uniform slip at V_{pl} (Fig. 10a),

$$\tau_0 = \tau_{ss}(V_{pl}) + \frac{\mu}{2c_s} V_{pl}. \quad (35)$$

The initial value of x was set as the steady-state value at $V = V_{pl}/2$, $x_{ss}(V_{pl}/2)$.

4 RESULTS OF EARTHQUAKE SEQUENCE SIMULATIONS

Calculated cumulative slip distributions are plotted in Fig. 11. Earthquakes were defined by a threshold of 0.1 m s^{-1} for the spatially maximum slip rate. Both models produced earthquakes on a regular basis, and the recurrence intervals of earthquakes were 157 yr and 81.3 yr in Model A and B, respectively. The ruptures were nucleated at around 7.7 to 8.5 km depth in Model A, and at around 7.7 to 9.9 km depth in Model B, and propagated in both upward and downward directions in both models. The downward extension of

the ruptures in which V exceeded 0.1 m s^{-1} were deeper in Model B ($z = 13.7 \text{ km}$) than in Model A ($z = 11.6 \text{ km}$). High pore pressure causes more frequent earthquakes with less slip per event and deeper penetration of ruptures. This is consistent with the expectation from the steady-state profiles (Fig. 10).

Before an earthquake, the peak shear stress at the d-nd transition was at depth of 9.66 km and 11.1 km for Model A and B, respectively (Fig. 12). It was slightly shallower than the steady-state case (Fig. 10a) because of a pre-seismic creep rate slower than V_{pl} . Earthquake ruptures propagated deeper than the peak shear stress, and were arrested with about a 1 to 1.5 km wide region of negative static stress drop within the ruptured area.

Interseismic spatio-temporal distributions of V and x are shown in Fig. 13. Note that x is linearly dependent on the void ratio e as $e = (x_0 - x)/(D - x_0)$. Two interseismic periods are shown in each panel, and the earthquakes indicated by yellow filling in Fig. 11 occurred in the middle of the plotted period, being recognized as horizontal discontinuities in the plots of V and sharp valleys in those of x . Interseismically, creeping motion invaded from both the shallow and deeper side of the locked region. The shallower creep motion is rarely observed on natural fault, and the condition for its occurrence was systematically studied by Kaneko *et al.* (2013). In the latter half of interseismic period, creeping regions produced aseismic transients. Earthquakes were nucleated at near the boundary of the locked region and the deep creeping region.

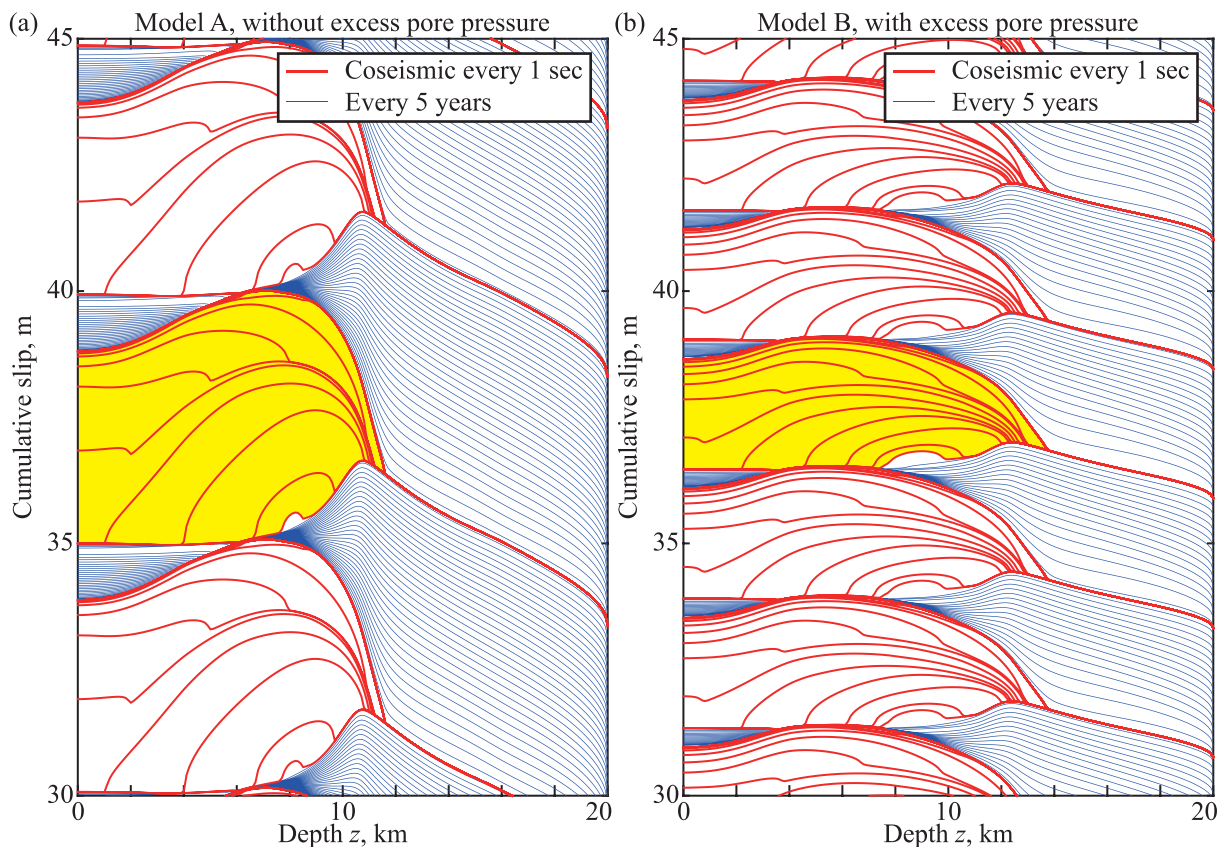


Figure 11. Cumulative slip distribution of (a) Model A (without excess pore pressure) and (b) Model B (with excess pore pressure) plotted in the same scale. Blue lines are plotted every 5 yr and red lines, every 1 s during earthquakes. The events with yellow filling are plotted in Figs 12 and 13.

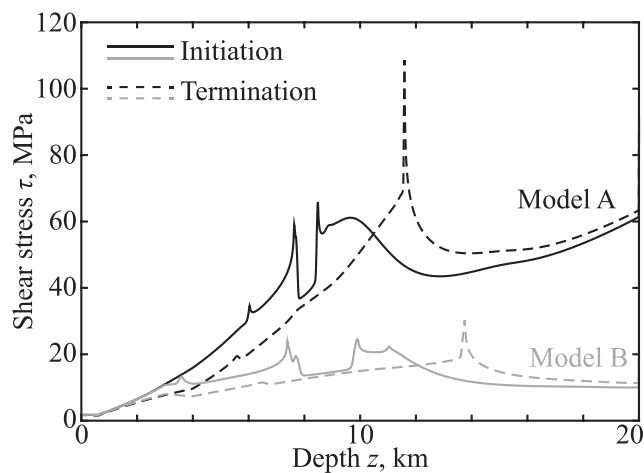


Figure 12. Shear stress distributions at initiation (solid lines) and termination (dashed lines) of earthquakes. The plotted events are indicated by yellow filling in Fig. 11.

After earthquakes, afterslip took place mainly below the ruptured area.

Rapid coseismic slip reduced the clast overlap zone width x , which increased due to time-dependent compaction by PSC more rapidly at a deeper part of the ruptured area because of higher T and σ_n^{eff} . Fault rocks below 6.2 km in Model A and below 7.3 km in Model B experienced significant compaction such that x/x_0 reached 0.99. The downward penetration of shallow creep truncated the time-dependent compaction (Fig. 13c). Acceleration of the

deep creep caused a decrease in x there, causing a non-monotonic depth-distribution of x . Aseismic transients in this region produced temporal oscillation in x in Model B (Fig. 13d). The earthquakes were nucleated within this region of decreased x (i.e. increased porosity).

5 DISCUSSION

5.1 A friction law based on a microphysical model

It was demonstrated in the above that the rather involved microphysical model proposed by H&S can be implemented in dynamic earthquake sequence simulations after modification to include a positive direct effect and an evolution effect. A great advantage of the existence of a microphysical model is that parameters and variables in it can be compared with fault zone structures. This means that the law can be scaled from laboratory experiments to natural faults with a concrete physical picture although the selection of the parameter values may not be straightforward due to idealization of the model.

In conventional RSF laws, the state variable θ represents recent slowness ($1/\text{slip rate}$) of the fault motion. In the aging law, for example, it is averaged slowness with a fading-memory weighting function (Ruina 1983). Although the mathematical definition of θ is clear, what it exactly correlates with in terms of fault zone microstructure is obscure. In the present friction law, x is the state variable which is related to interface porosity or void ratio given by $(x_0 - x)/(D - x_0)$. Therefore, the spatio-temporal distribution of the state variable x (lower row of Fig. 13) may be able to be coupled

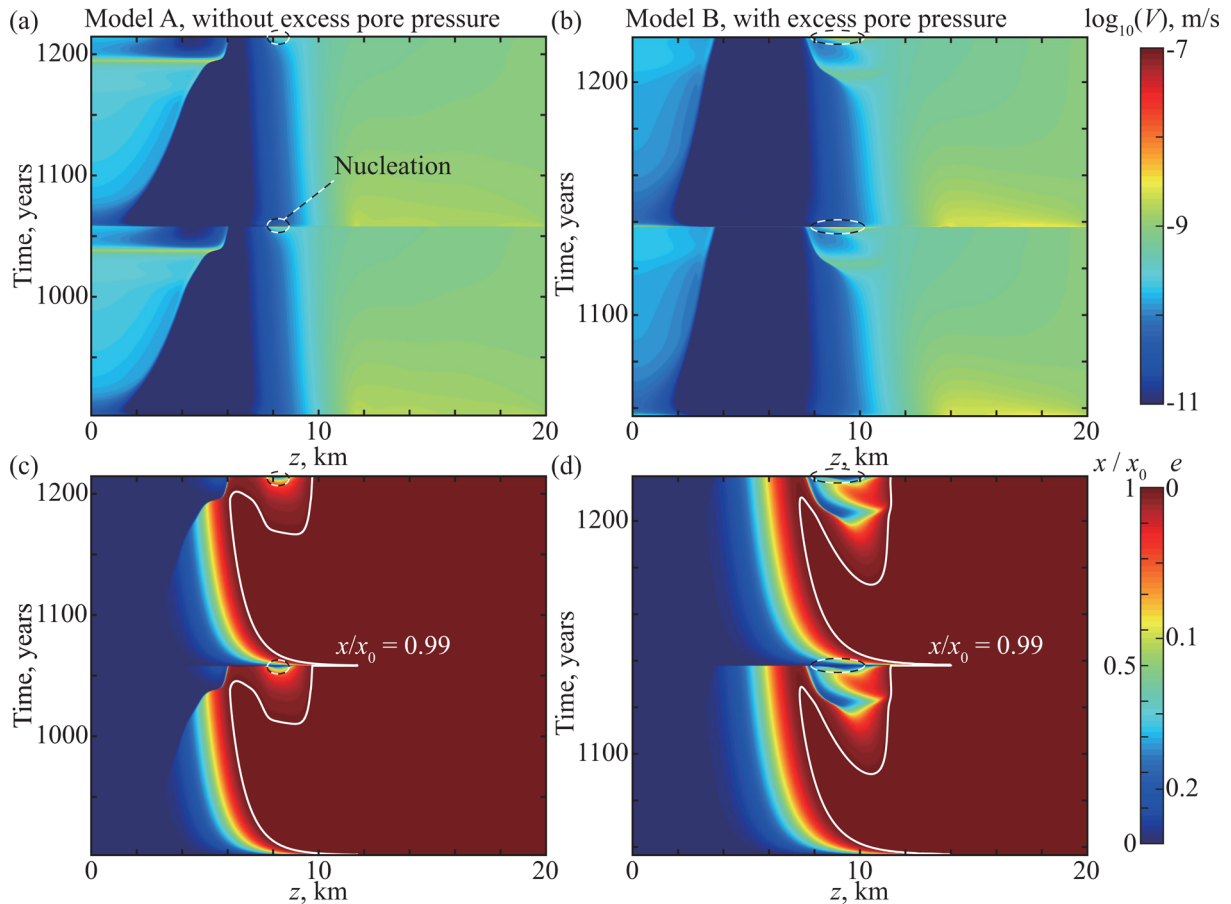


Figure 13. Spatio-temporal distributions of interseismic V (upper) and x or void ratio e (lower) for Model A (left) and B (right). The earthquakes indicated by yellow filling in Fig. 11 occurred in the middle of the plotted period. Black-white dashed ellipses indicate locations of earthquake nucleation. Note that x decreases (i.e. porosity increases) during aseismic transients and before nucleation.

with other physical processes such as pore pressure fluctuation and fluid flow.

In the present simulations, the parameters (Table 3) are selected as characteristic values for an ultracataclasite layer with a thickness of 1 cm. This works for the purpose of demonstration of the friction law by H&S and of qualitative discussion, but deserves further examination. Evidence of pressure solution in much wider zones of foliated cataclasite and phyllonite [e.g. about 10 m in Median Tectonic Line fault zone exposed in Miyamae, Mie prefecture Japan (Jefferies *et al.* 2006)] may suggest a much larger effective value of w . On the other hand, a very thin submillimetre thick principle slip plane is often recognized within or at the boundary of an ultracataclasite layer (e.g. Chester & Chester 1998; Heermance *et al.* 2003; Rice 2006). One way to account for such a complex fault zone structure would be connecting elements with different properties (e.g. D , w and f_0). This deserves further investigation although the friction law would become much more complicated.

The microphysical model by H&S successfully described the d-nd transition, but there are limitations and further development deserves future study. First, the applicable range of f_0 is restricted from $\pi/8$ to $\pi/4$, as noted by H&S. At lower f_0 , the clast overlap zone disappears. At higher f_0 , the approximation to cylindrical clasts is no longer valid, and an approximation to other shape (e.g. lenticular ‘fish’ Bos & Spiers 2002b) may be more appropriate which would yield a different geometrical factor and dependency on x compared with the present model. Second, the current model by H&S cannot express the evolution effect well inside the

dilatant regime. To account for this, a modification was applied in this study by introducing the b -term to the friction of wavy foliation for practical purposes, but a microphysically based modification would be more suitable. Third, connections to different deformation regimes (e.g. crystalline plasticity at higher temperatures, and dynamic weakening at coseismically high slip rates) are important as discussed in following sections. This issue was partly studied using phenomenological conventional RSF laws (e.g. Noda & Lapusta 2010, 2013; Shimamoto & Noda 2014), but the existence of the microphysical model and extension of it presumably help us build up more concrete understanding. Fourthly, evolution of fault structures other than x should be incorporated towards more realistic modelling. Localization of strain rate and reduction in w may be the case under conditions that the friction law is rate-weakening. Grain size distribution and its heterogeneity across a fault zone may evolve by grain crushing and solution-precipitation processes. Establishing a fault model which yields realistic fault zone structure is an important future challenge.

5.2 Strong barrier beneath the seismogenic zone

One of the most significant differences of the presented models from those simulated using logarithmic RSF laws is the existence of the strong barrier directly beneath the seismogenic zone. As an example, one of the models by Lapusta *et al.* (2000) is shown in Fig. 14. In this case, the critical wavelength λ_{cr} ($=4h_*$ where h_*

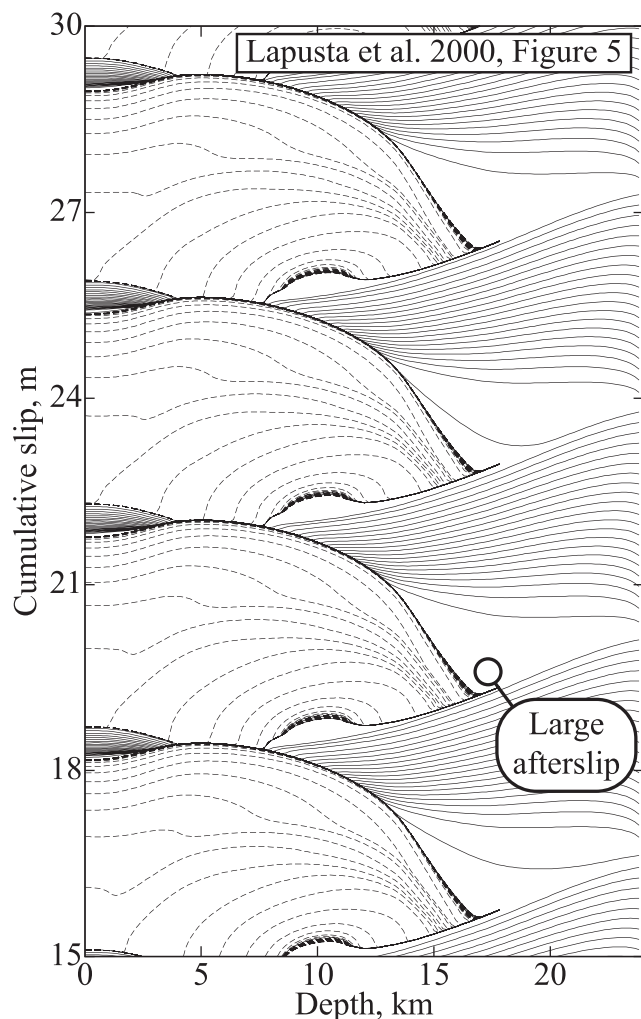


Figure 14. One of the models of earthquake sequences using a logarithmic RSF law (the aging law) modified from fig. 5 in Lapusta *et al.* (2000). The plate rate V_{pl} is 35 mm s^{-1} ($= 1.110 \times 10^{-9} \text{ m s}^{-1}$). In the seismogenic zone, $\sigma_e^{\text{eff}} = 50 \text{ MPa}$, steady-state rate dependency $a - b = 0.004$. The condition is comparable to Model B in this study. The solid lines represent slip distribution every 5 yr, and the dashed lines are plotted during earthquakes every 1 s. The time intervals of the lines are identical to Fig. 11.

is the critical cell size in a spectral basis) in the seismogenic zone is 3.76 km, and the steady-state rate dependency $d\tau_{ss}/d\ln(V)$ is -0.2 MPa from 4 to 13.5 km and increases to about 1 MPa around 20 km. In Model B, λ_{cr} is in a similar range to that of Lapusta *et al.* (2000) within the seismogenic zone except in a locally unstable region at the onset of the dilatant deformation (Fig. 10d), but the rate dependency $d\tau_{ss}/d\ln(V)$ at $V = V_{pl}$ reaches about 10 MPa (Fig. 10b). This large peak is because of the strong rate-strengthening effect of PSC which is linearly viscous. It causes significantly smaller amounts of afterslip. In Fig. 14, it is obvious that slip accumulation around the arrested rupture front accelerates substantially after the earthquakes. On the contrary, such acceleration is minor in the present models (Fig. 11).

The frictional parameters adopted by Lapusta *et al.* (2000) are based on friction experiments of granite gouge at hydrothermal conditions (Blanpied *et al.* 1998) under which PSC most likely operates (Blanpied *et al.* 1995). The experimental data show an increase in the rate dependency as T increases from about 200–300 °C. This

is qualitatively similar to the friction law in this study, but less remarkable. The granite gouge may have different characteristics from the clay–quartz mixture studied by H&S or its analogue material (e.g. Bos *et al.* 2000a) because of much smaller clay content and existence of different phases such as plagioclase. Even if the linearized characteristics (e.g. $A - B$) were similar, the difference in the functional form (logarithmic versus linear) should cause a significant difference in afterslip behaviour. Fig. 12 shows that the stress increase near the arrested rupture front is smaller than, but a good fraction of, the absolute value of the pre-stress. This indicates that the slip rate would increase by a smaller factor than 2 for ideally linearly viscous cases. In the logarithmic RSF laws, the rate dependency parameters a and b are typically much smaller than the friction coefficient, and thus stress perturbation much smaller than the absolute shear stress may cause a change in the slip rate by orders of magnitude.

Shimamoto & Noda (2014) conducted similar earthquake sequence simulations accounting for a transition to power-law creep. Their simulations also have a major peak in the rate dependency beneath a seismogenic zone (fig. 3b in Shimamoto & Noda 2014), and showed earthquake sequences with smaller afterslip for smaller power exponent (fig. 10 in Shimamoto & Noda 2014). Investigation of characteristics of natural afterslip (e.g. amplitude and decay rate) would be important in understanding the dominant deformation mechanism directly below the seismogenic zone.

It should be pointed out that the absolute value of frictional resistance is not an important factor affecting the afterslip behaviour. At the d–nd transition, PSC becomes active at the deeper region due to larger T and σ_n^{eff} , decreasing frictional resistance and causing a peak in the distribution of τ . This weakened part was dragged by shallower coseismic slip by much less than in the case of a logarithmic RSF because of remarkable rate-strengthening owing to linearly viscous PSC. The onset of PSC weakens the fault for clay–quartz mixture in terms of absolute stress level, but strengthens it in terms of response to stress perturbation.

5.3 Transition to crystalline plasticity and need for dynamic weakening

In this study, the focus is placed on the transition between dilatant deformation, which may be approximated by a logarithmic RSF law, to non-dilatant deformation caused by efficiently active PSC (d–nd transition). At a greater depth, a transition to crystalline plasticity (brittle–plastic transition, BPT) is expected (e.g. Bos & Spiers 2002b, among others). Geological observations suggest that earthquake ruptures may reach the BPT, which produces repeated overprint of coseismic deformation generating pseudo-tachylytes and interseismic ductile deformation generating mylonites (e.g. Sibson 1980; Hobbs *et al.* 1986; Lin *et al.* 2005). On the other hand, in the present simulations, the earthquake ruptures were arrested near the d–nd transition, and could not propagate spontaneously further down.

Spontaneous rupture propagation requires dynamic stress drop. It is sometimes pointed out that a fault is weakened by PSC as realized in the present models below the d–nd transition. If so, even smaller frictional resistance at coseismic slip rates is required there for spontaneous rupture propagation. τ_{ss} as a function V predicted by the present friction law (Fig. 7a for example) suggests that it is very difficult for a region of pre-seismic non-dilatant deformation to produce stress drop. The exercise demonstrated in this paper has explicitly shown that implementation of dynamic

weakening (e.g. Rice 2006) at coseismic slip rate is essential for more realistic modelling of earthquake sequences in which dynamic ruptures reach a regime of crystalline plasticity or the transition to it.

5.4 Comparison with geophysical observations

In the deeper extent of plate boundary faults than the seismogenic zone, slow earthquakes and slow slip events were observed (e.g. Hirose *et al.* 1999; Obara 2002, among many). Episodic tremor and slip events propagate in a parabolic manner (Ide 2010), which can be explained by a linearly viscous fault constitutive law (Ando *et al.* 2012). Although diffusion of pore fluid along a fault may also explain the parabolic migration, the linearly viscous nature of PSC is another candidate. Quantitative examination would be an important future study.

There are some studies reporting rate dependency of a fault obtained from geodetic observation of post-seismic deformation on a strike-slip fault. But the behaviour of a deep part of the fault is difficult to capture because stations close to the fault trace are dominated by the behaviour of a shallow part of the fault. Johnson *et al.* (2006) and Barbot *et al.* (2009) studied post-seismic deformation of the Mw 6.0 2004 Parkfield earthquake, and reported the steady-state rate dependency of 5–50 kPa and 0.7 MPa, respectively. Later, Chang *et al.* (2013) analysed the same data set, while allowing depth-variable parameters. They tested different values for the depth-gradient of the deviatoric stress in order to estimate the pre-stress, and reported rate dependencies beneath the coseismic slip. The steady-state rate dependency was below 0.1 MPa for the smallest stress gradient 5 MPa km⁻¹, and on the order of MPa for the other cases. At 12 km depth, the majority of their depth-dependent models show 2–4 MPa of rate dependency.

Model A, which assumed no excess pore pressure at depth, clearly contradicts with the example of the 2004 Park field earthquake. In Model B, in which $\sigma_n^{\text{eff}} \leq 50$ MPa, the maximum rate dependency is larger by a factor 3, which is still significantly too large. Detailed parameter study may deserve further study, but improvement of the microphysical model and the friction law is probably more important. For example, it may be the case that the d-nd transition predicted by the model by H&S is too sharp. Niemeijer & Spiers (2005) considered a similarly sharp transition, and pointed out that it contradicts their experimental data for a muscovite-halite mixture. They also pointed out that grain size distribution, instead of single-sized clasts, smoothed out the transition and yielded much milder rate-strengthening behaviour. The difference by a factor of 3 may be explained by this effect. Another example is the effect of dynamic weakening at coseismic slip rate. It probably promotes rupture propagation further down by generating additional dynamic stress drop, and ruptures would be arrested at a greater depth than in the present model. Then the rate dependency just ahead of the arrested rupture may be much smaller than that near the d-nd transition in the steady-state picture.

6 CONCLUSIONS

A microphysical model for a friction law for clay–quartz mixture accounting for PSC of quartz at low slip rates and onset of dilatant deformation (d-nd transition) at high slip rates (den Hartog & Spiers 2014) was implemented in dynamic earthquake sequence simulations (Lapusta *et al.* 2000) after modification. The model was a version of RSF law, but frictional elements of clay had a constant

friction coefficient, causing zero instantaneous rate dependency in the dilatant regime. Introduction of instantaneous rate dependency of clay, which is ubiquitously observed, gave the law a positive direct effect, which is important in considering coupling with the elastic surrounding medium (Rice *et al.* 2001).

The characteristics of the present friction law were investigated by numerically solving the steady-state, linearization around it and velocity steps. The power exponent of the porosity dependency of PSC compaction rate $n < 1$ causes a vanishingly small state-evolution slip distance d_c and state-evolution timescale d_c/V as V decreases. By assuming $n = 1$, the timescale d_c/V stayed finite as V decreased, and earthquake sequence simulations became affordable in terms of computational cost. The increase in n causes milder enhancement of rate-weakening at the d-nd transition. Velocity steps across the d-nd transition showed qualitatively asymmetric behaviour (peak-decay for a positive step, and almost monotonic decay for a negative step).

2-D dynamic earthquake sequence simulations for a strike-slip fault (Lapusta *et al.* 2000) were demonstrated. Two models with and without excess pore pressure at depth were tested. The earthquakes were nucleated in the region of enhanced rate-weakening near the d-nd transition, and arrested at a point a few kilometres deeper than the peak pre-stress. Higher pore pressure caused a wider seismogenic zone, deeper rupture propagation, smaller slip per event and a shorter recurrence interval. One of the most significant differences from models with conventional logarithmic RSF laws was the much smaller amount of afterslip near the arrested rupture front at depth. The deformation conditions directly below the d-nd transition are associated with linear rate dependency that is much stronger than the logarithmic one. Observation of afterslip may give a key to understand dominant deformation mechanism directly below the seismogenic zone.

The simulated ruptures could not propagate spontaneously in a region of pre-seismic non-dilatant deformation. This is because the region cannot produce a positive stress drop, owing to weakness of the fault against slow deformation. Geological observations (e.g. Sibson 1980) evidenced that at least some of the ruptures reached the brittle-plastic transition. In order to realize such a model, it is probably required to reduce the frictional resistance at coseismic slip rates even below the level of the resistance against slow deformation in the regime of active PSC.

A friction law derived from a microphysical model has a great advantage over phenomenological laws in that the parameters and variables in it can be compared with fault zone structures, but idealization of complex fault zone structures is not straightforward. For example, the width of the shearing zone may change with time, and so do effective structural parameters (e.g. grain size, quartz fraction and so on). Study of earthquake sequences based on a microphysical model is a key to physical understanding of earthquake-related phenomena, and further investigation is urged including revision of the microphysical model through interdisciplinary collaboration between experimental, theoretical, numerical and observational studies.

ACKNOWLEDGEMENTS

This study is supported by MEXT/JSPS KAKENHI Grant Numbers 26109007 and 26870918. I would like to express my sincere gratitude to Sabine den Hartog for discussion and English check before submission. I gratefully appreciate comments by Brett Carpenter, Chris Spiers, Miki Takahashi and André Niemeijer which were very useful in improving the manuscript.

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APPENDIX A: EXPRESSIONS OF FACTORS IN THE FRICTION LAW

The expressions of the factors appearing in the friction law are summarized here for completeness. They are derived by den Hartog (2013) and presented in den Hartog & Spiers (2014).

Clast body zone

$$F_B(x) = 1 - \frac{1}{L(D-2x)} \left(\frac{\pi D^2}{4} - \frac{16x^2(D-x) + 3x^3}{6\sqrt{Dx-x^2}} \right) \quad (A1)$$

$$\eta_B(x) = \frac{RT}{A_{HS}k_+\Omega^2} \frac{D}{n_{cell}L(D-2x)^2} \times \left(\frac{\pi D^2}{4} - \frac{16x^2(D-x) + 3x^3}{6\sqrt{Dx-x^2}} \right) \quad (A2)$$

Clast overlap zone

$$F_O(x) = 1 - \frac{2\sqrt{Dx-x^2}}{L} \quad (A3)$$

$$\eta_O(x) = \frac{RT}{k_+\Omega^2} \frac{x}{n_{cell}L} \quad (A4)$$

Dilatant deformation

$$F_{dil}(x, \mu_{dil}) = \frac{L^2 + x^2}{L^2 - (\mu_{dil}x)^2} \quad (A5)$$

State-evolution law

$$F_x = \frac{\pi D}{n_{cell}L} \quad (A6)$$

$$\eta_{xc} = \frac{2RTL}{\pi k_+\Omega^2 D} \quad (A7)$$

A_{HS} is a geometrical factor from 2 to π and $A_{HS} = 2$ is adopted in this study, R is the gas constant $8.314462175 \text{ J mol}^{-1} \text{ K}^{-1}$, Ω is molar volume of quartz $2.27 \times 10^{-5} \text{ m}^3 \text{ mol}^{-1}$ and k_+ is a dissolution rate coefficient expressed as (Tester *et al.* 1994)

$$k_+ = 276 \exp \left(-\frac{90.1 [\text{kJ mol}^{-1}]}{RT} \right) [\text{mol m}^{-2} \text{ s}^{-1}]. \quad (A8)$$