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- Microphysical model developed for frictional shear of illite-quartz gouge
- Model predicts main trends observed in experiments
- Velocity-weakening (seismogenic) slip explained by pressure solution of quartz

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A microphysical model for fault gouge friction applied to subduction megathrusts

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Abstract A microphysical model is developed for the steady state frictional behavior of illite-quartz fault gouge and applied to subduction megathrust P - T conditions. The model assumes a foliated, phyllosilicate-supported microstructure which shears by rate-independent frictional slip on the aligned phyllosilicates plus thermally activated deformation of the intervening quartz clasts. At low slip rates or high temperatures, the deformation of the clasts is easy, accommodating slip on the foliation without dilatation. With increasing velocity or decreasing temperature, the shear of the clasts becomes more difficult, increasing bulk shear strength, until slip is activated on inclined portions of the phyllosilicate foliation, where it anastomoses around the clasts. Slip at these sites leads to dilation involving clast/matrix debonding, balanced, at steady state, by compaction through thermally activated clast deformation. Model predictions, taking pressure solution as the thermally activated mechanism, show three regimes of velocity-dependent frictional behavior at temperatures in the range of 200–500°C, with velocity weakening occurring at 300–400°C, in broad agreement with previous experiments on illite-quartz gouge. Effects of slip rate, normal stress, and quartz fraction predicted by the model also resemble those seen experimentally. Extrapolation of the model to earthquake nucleation slip rates successfully predicts the onset of velocity-weakening behavior at the updip seismogenic limit on subduction megathrusts. The model further implies that the onset of seismogenesis is controlled by the thermally activated initiation of fault rock compaction through pressure solution of quartz, which counteracts dilatation due to slip on the fault rock foliation.

1. Introduction

Numerical modeling of subduction zone seismogenesis, and of slow slip events [e.g., Liu and Rice, 2005, 2007, 2009; Matsuzawa et al., 2010; Noda and Lapusta, 2013; Perfettini and Ampuero, 2008; Yang et al., 2012], relies on input data describing the frictional properties of megathrust fault rocks under in situ conditions. However, the relevant pressures (P), temperatures (T), slip velocities (V), and shear strains (γ) are seldom approached in friction experiments on compositionally representative materials [e.g., Moore et al., 1986a, 1986b, 1989; Morrow et al., 1992; Saffer and Marone, 2003].

Recently, Den Hartog et al. [2012a, 2013] reported the first large strain ($\gamma \leq 170$) friction experiments on realistic megathrust gouge compositions under in situ P - T conditions. Simulated gouges composed of 65:35 illite:quartz and muscovite:quartz were used to capture the progressive metamorphic evolution of the pelitic fault rocks expected at seismogenic depths within subduction megathrusts (i.e., at ~5–40 km, ~150–350°C [Hyndman et al., 1997]), assuming illite to be stable at ~150–200°C with muscovite replacing it above 200–300°C [e.g., Van de Kamp, 2008]. The experiments were mostly conducted at an effective normal stress (σ_n^{eff}) of 170 MPa, a pore fluid pressure (P_f) of 100 MPa, and sliding velocities (V) of 1–100 $\mu\text{m/s}$, investigating temperatures in the range of 100–600°C. In the terminology of Rate and State Friction (RSF) [Dieterich, 1978, 1979; Ruina, 1983], both gouges showed stable, velocity-strengthening (or near-neutral) behavior at low temperatures (Regime 1), potentially unstable, velocity weakening at intermediate temperatures (Regime 2) and velocity strengthening at the highest temperatures investigated (Regime 3) [Den Hartog et al., 2012a, 2013]. For illite-quartz gouge (Figure 1), velocity weakening (Regime 2) was observed between 250°C and 400°C, whereas for muscovite-quartz gouge it occurred at 350–500°C. The observed three-regime behavior resembles that reported for granitic and gabbroic gouges [Blanpied et al., 1991, 1995, 1998; He et al., 2007], with the combined velocity-weakening regimes of the two gouges offering a possible explanation for the seismogenic zone within subduction megathrusts [Den Hartog et al., 2013].

Figure 1. Synoptic diagram summarizing the key results obtained by Den Hartog and Spiers [2013; see also Den Hartog et al., 2012a] from ring shear experiments on illite-quartz gouges, at near in situ megathrust conditions. (a) Rate and state friction parameter ($a-b$) versus temperature for the sliding velocity steps shown. Data obtained at reference conditions of $\sigma_n^{\text{eff}} = 170$ MPa, $P_f = 100$ MPa using gouge with initial composition of 65:35 illite:quartz. Open symbols are calculated using peak friction values of stick-slip events and differ from the ($a-b$) values for stick-slip events originally reported by Den Hartog et al. [2012a] and Den Hartog and Spiers [2013], which were based on mean friction values. (b) Main trends observed in friction coefficient versus temperature (T), effective normal stress (σ_n^{eff}), and quartz fraction (f_{qtz}). The friction coefficient is defined as the ratio of shear stress to effective normal stress and neglects cohesion. All curves represent μ at the apparent yield point, except the black dotted curve, which gives μ versus temperature at a shear strain of 50. The curve showing the effect of σ_n^{eff} at low temperatures (i.e., falling in Regime 1 of Figure 1a) is dashed at low effective normal stresses where the sample cohesion is significant. (c–e) Schematic representation of the displacement of the ($a-b$) versus temperature curve shown in Figure 1a (black curve) as a result of changes in: (Figure 1c) slip rate, (Figure 1d) effective normal stress, and (Figure 1e) quartz content. The question mark in Figure 1d indicates uncertainty in the interpretation of the effect of quartz content on the ($a-b$) versus temperature profile; the experimental data could also be explained by vertical depression of the curve.

Such an interpretation is supported by the fact that illite-quartz gouge shows a shift of Regime 2 behavior toward lower temperatures, with decreasing sliding velocity (Figures 1a and 1c), moving toward the seismogenic window of 150–350°C as rupture nucleation velocities are approached ($\sim 10^{-9}$ m/s [e.g., Segall and Rice, 2006]). This shift, along with the effects of quartz content and effective normal stress reported by Den Hartog and Spiers [2013] (Figure 1), offers a means of extrapolating the frictional behavior seen in experiments to true in situ megathrust conditions. However, the observed trends are only poorly constrained, highlighting the need for a mechanism-based model.

Previously, we explained the behavior of illite- and muscovite-quartz gouges qualitatively, in the framework of the microphysical model for steady state shearing of halite-phyllsilicate and quartz-phyllsilicate fault gouges proposed by Niemeijer and Spiers [2005, 2006, 2007; see also Bos and Spiers, 2001, 2002]. This model assumes shear by thermally activated deformation of the halite or quartz clasts, by pressure solution, plus (dilatant) intergranular slip on the subordinate phyllsilicate phase. However, the model is not appropriate for quantitative comparison with experiments on phyllsilicate-rich gouges. This is because it assumes that the clasts are separated by only a thin film of phyllsilicates, as opposed to a matrix-supported microstructure. In addition, the dominant, thermally activated deformation mechanism affecting the quartz clasts in our experiments was inferred to be Stress Corrosion Cracking (SCC), with pressure solution likely playing a lesser role [Den Hartog et al., 2012a, 2013]. The relative importance of these mechanisms could not be quantified, however.

In this paper, we derive a quantitative model for the steady state frictional behavior of matrix-supported phyllsilicate-quartz gouges deforming by frictional slip on the phyllsilicate foliation plus thermally activated deformation of the quartz clasts. The approach resembles that employed by Niemeijer and Spiers [2007] for clast-supported gouge, but avoids several unnecessary assumptions. We base our model on the

Table 1. Explanation of Symbols Used in the Current Paper

Symbol	Meaning
a	Magnitude of direct effect (RSF) (-)
A	Shape factor (-)
A_{pore}	Pore area available for quartz-on-quartz precipitation (m^2)
A_{pore-c}	Pore area per clast at critical state (m^2)
A_{qtz-b}	Average horizontal area occupied by a single clast body (m^2)
A_{qtz-o}	Area over which an overlap displaces during slip (m^2)
A_x	Horizontal position along phyllosilicate foliation (m)
A'_{seg}	Area of clast segment in overlap zone, perpendicular to shear plane (m^2)
b	Magnitude of evolution effect (RSF) (-)
D	Clast diameter and length (m)
f_{qtz}	Volume fraction of quartz (-)
I	$k_+ \Omega$ (ms^{-1})
k_+	Dissolution rate coefficient for quartz ($mol\ m^{-2}\ s^{-1}$)
k_f	Constant depending on the 3-D clast shape, 1/4 for cylinder assumed here (-)
L	Horizontal clast spacing, unit cell length (m)
L_1	Length of unit cell over which the anastomosing foliation displays a positive inclination (m)
L_2	Length of unit cell over which the anastomosing foliation displays a negative inclination (m)
ℓ_1	(Length of) line approximating anastomosing foliation with positive inclination (m)
ℓ_2	(Length of) line approximating anastomosing foliation with negative inclination (m)
n	Exponent in relation describing pore area, pore shape evolution parameter (-)
R	Gas constant ($J\ mol^{-1}\ K^{-1}$)
s	Vertical clast spacing, unit cell height (m)
S_{n1}	Normal stress component on ℓ_1 (Pa)
S_{s1}	Shear stress component on ℓ_1 (Pa)
T	Temperature (K)
V	Sliding velocity (ms^{-1})
w	Average sample thickness (m)
x	Overlap of adjacent horizontal rows of clasts (m)
x_0	Overlap of adjacent horizontal rows of clasts in the undilated (dense) state (m)
$\dot{\epsilon}_{comp}$	Rate of compaction by a thermally activated mechanism (s^{-1})
$\dot{\epsilon}_{dil}$	(Dilational) strain normal to anastomosing phyllosilicates (-)
$\dot{\epsilon}_{dil}$	Rate of dilatation due to slip on anastomosing phyllosilicates (s^{-1})
$\dot{\epsilon}_m$	Macroscopic rate of compaction (s^{-1})
ϕ	Porosity (-)
ϕ_c	Porosity at critical state (-)
$\dot{\gamma}$	Shear strain rate (s^{-1})
$\dot{\gamma}_B$	Shear strain rate due to slip on the horizontal foliation and clast body deformation (s^{-1})
$\dot{\gamma}_{dil}$	(Shear) strain parallel to anastomosing phyllosilicates (-)
$\dot{\gamma}_{dil}$	Shear strain rate due to dilatant slip on anastomosing phyllosilicates (s^{-1})
$\dot{\gamma}_m$	Macroscopic or imposed shear strain rate (s^{-1})
$\dot{\gamma}_{non-dil}$	Shear strain rate due to non-dilatant deformation (s^{-1})
$\dot{\gamma}_O$	Shear strain rate due to slip on the horizontal foliation and clast overlap deformation (s^{-1})
$\dot{\gamma}_{qtz-b}$	Shear strain rate due to thermally activated deformation of quartz clast bodies (s^{-1})
$\dot{\gamma}_{qtz-o}$	Shear strain rate due to thermally activated deformation of quartz clast overlaps (s^{-1})
λ	Width at the base of the clast overlap (m)
μ	Macroscopic or measured friction coefficient (-)
μ	Friction coefficient within phyllosilicates (-)
σ_n^{eff}	Macroscopic effective normal stress (Pa)
τ_1	Portion of macroscopic shear stress that acts over distance L_1 (Pa)
τ_2	Portion of macroscopic shear stress that acts over distance L_2 (Pa)
τ_B	Shear stress due to slip on the horizontal foliation and clast body deformation (Pa)
τ_{dil}	Shear stress to activate/maintain slip on anastomosing foliation (Pa)
τ_m	Macroscopic shear stress (Pa)
τ_O	Shear stress due to slip on the horizontal foliation and clast overlap deformation (Pa)
τ_{ph}	Shear stress associated with frictional slip on phyllosilicates (Pa)
τ_{qtz-b}	Shear stress associated with deformation of quartz clast bodies (Pa)
τ_{qtz-o}	Shear stress associated with deformation of quartz clast overlaps (Pa)
Ω	Molar volume of quartz ($m^3\ mol^{-1}$), $2.27 \times 10^{-5}\ m^3\ mol^{-1}$
Ψ_{dil}	Dilatation angle (deg)
Ψ_{fr}	Mean inclination of anastomosing foliation (deg)

microstructures developed in our previous illite-quartz experiments. Since almost nothing is known about the deformation of quartz clasts by SCC under the relevant (P - T) conditions, we focus on pressure solution, with the aim of evaluating to what extent this process, combined with friction on the phyllosilicates, can explain the behavior seen in our experiments. Favorable agreement is obtained, suggesting that pressure solution was more important in our experiments than we previously inferred. We go on to apply our model

Figure 2. Representative microstructures seen in illite-quartz gouges deformed at (near) in situ megathrust P - T conditions, i.e., $P_f = 100$ MPa and various values of T , σ_n^{eff} , and V . All samples shown are characterized by negative to near-neutral (a – b) values, with the possible exception of the sample shown in Figure 2f (refer Figure 1). The experimentally imposed shear sense (dextral) is indicated in the top right corner of each figure. The foliations visible in some of the micrographs represent R_1 or P shear band foliations and therefore need not be parallel to the imposed shear direction. Note that pores are visible as black patches in SEM micrographs (Figures 2a–2d), whereas pores and/or thin areas are white in TEM images (Figures 2e and 2f). (a) Microstructure after deformation at $T = 250^\circ\text{C}$, $\sigma_n^{eff} = 170$ MPa, and $V = 1$ – 100 $\mu\text{m/s}$, obtained using Focused-Ion Beam Scanning Electron Microscopy (FIB-SEM). (b) FIB-SEM microstructure after deformation at $T = 300^\circ\text{C}$, $\sigma_n^{eff} = 170$ MPa, and $V = 1$ – 100 $\mu\text{m/s}$. (c) FIB-SEM microstructure after deformation at $T = 350^\circ\text{C}$, $\sigma_n^{eff} = 170$ MPa, and $V = 10$ $\mu\text{m/s}$. (d) FIB-SEM microstructure after deformation at $T = 500^\circ\text{C}$, $\sigma_n^{eff} = 50$ MPa, and $V = 1$ – 100 $\mu\text{m/s}$. (e) Microstructure after deformation at $T = 350^\circ\text{C}$, $\sigma_n^{eff} = 170$ MPa, and $V = 10$ $\mu\text{m/s}$, obtained using Transmission Electron Microscopy (TEM). (f) TEM microstructure after deformation at $T = 500^\circ\text{C}$, $\sigma_n^{eff} = 170$ MPa, and $V = 10$ $\mu\text{m/s}$.

to predict the temperature range in which velocity weakening occurs at in situ nucleation velocities on subduction zone megathrusts.

2. Microstructural Model

We start by summarizing the microstructures developed in our experiments on illite-quartz gouges [Den Hartog *et al.*, 2012a, 2012b]. These are used as a basis for defining an idealized microstructure and the deformation mechanisms to be modeled. For the symbols used, see Table 1.

2.1. Microstructural Observations

Illite-quartz gouges sheared at (near) in situ megathrust P - T conditions show a relatively uniform microstructure composed of equiaxed, micron-sized quartz clasts distributed throughout a finer phyllosilicate matrix [Den Hartog *et al.*, 2012a, 2012b] (Figure 2). The phyllosilicates are frequently aligned (Figures 2a, 2d, 2e, and 2f), forming a foliation that anastomoses around the quartz clasts (Figures 2d and 2f) and is often parallel to the P shear orientation of Logan *et al.* [1992] (Figures 2a, 2d, and 2f) or to the R_1 or Y orientation (Figure 2e). Elongate pores accounting for a porosity of several percent are developed preferentially at clast/matrix interfaces oriented favorably for extension and hence for debonding—i.e., at a high angle to the local incremental extension direction (Figures 2a–2d). No evidence was visible for the deformation of individual clasts by pressure solution (e.g., tailed porphyroclasts). Rather, the blocky to sub-angular or rounded shape of the quartz clasts, plus the reduction in clast size compared to the starting material (0.5–5 μm [Den Hartog *et al.*, 2012a]), notably at high temperature (Figures 2d and 2f), suggests deformation, or at least a phase of grain size reduction, by cataclastic processes such as SCC. However, this does not rule out small compactional or shear strains by pressure solution. Indeed, the oval shape of the finest clasts (Figure 2f) may point to deformation by pressure solution. Clearly, key features of the microstructure are phyllosilicate matrix support, and dilatant porosity at extensional quartz-phyllosilicate interfaces.

Figure 3. Assumed model microstructure for illite-quartz gouge undergoing shear deformation. (a) Representative volume of the basic shear band microstructure, showing cylindrical quartz clasts embedded in a matrix of aligned phyllosilicates. The phyllosilicates define both horizontal and curved foliations, the latter anastomosing around the quartz clasts. The anastomosing foliation makes tangential contact with the upper and lower extremities of each clast, but is drawn here with an exaggerated amplitude for clarity. Note the porosity (clast-matrix debonding) developed at the quartz-phyllosilicate interfaces undergoing incremental extension. The main microstructural elements, and related deformation processes, are indicated. Type B zones indicate horizontal regions containing the quartz clast bodies surrounded by a horizontal phyllosilicate foliation. These zones are characterized by slip on the horizontal phyllosilicates and simple shear of the clast bodies. Type O zones contain anastomosing phyllosilicates and the overlapping or interlocking extremities of the clasts (clast overlaps). These deform by slip on the horizontal phyllosilicates at the zone margins (the interface between the O and B zones) plus shear displacement of the clast overlaps as shown, or, if shear stresses become high enough, by slip on the anastomosing foliation and accompanying clast-matrix debonding. (b) Model unit cell and associated microstructural state variables. The unit cell dimension perpendicular to the plane of view is D .

2.2. Idealized Microstructure and Assumed Deformation Processes

From the above, we infer that shear deformation involved slip on internally foliated P, R_1 , and Y shear bands. Deformation within these bands appears to have featured sliding on the phyllosilicate foliation accommodated by (i) deformation of the intervening quartz clasts by processes such as SCC and/or pressure solution, or (ii) dilatation at clast-matrix interfaces subject to extension. We assume that the strength of the samples was controlled by the mechanical behavior of the active shear bands. To model sample strength, we will focus on Y shears, noting that the simple angular relationship between Y, P, and R_1 shears means that overall gouge behavior will be insensitive to the dominant shear band type. We assume that large strain shear deformation must be

Figure 4. Mechanical analogue diagram showing the assumed deformation processes in the model microstructure of Figure 3. Elements 1 and 2 show the processes that are active in Type B zones and at the margins of Type O zones, respectively, i.e., slip on the horizontal phyllosilicate foliation (elements 1a and 2a), accommodated by serial, thermally activated deformation of the quartz clast bodies (element 1b) and overlaps (element 2b). Element 3 represents dilatant slip on the anastomosing phyllosilicate foliation within Type O zones. Dilatant slip in the O zones results only in displacement of the (vertically) adjacent B zones with respect to each other. Note that the anastomosing foliation in the bottom diagrams is drawn with an exaggerated amplitude for clarity (cf. Figure 3). The variables are defined in the text.

accommodated by a geometrically suitable combination of R_1 , P , and/or Y shears and that slip on these systems controls strength.

Our idealized model for the steady state microstructure of illite-quartz gouge deforming by horizontal (Y) shear is shown in Figure 3. The model is quasi two dimensional in that it consists of cylindrical quartz clasts of diameter and length D embedded in a phyllosilicate matrix. The clasts are uniformly distributed in a regular packing characterized by a constant horizontal clast spacing L and a vertical spacing s , i.e., by the unit cell shown in Figure 3. Adjacent horizontal rows of clasts overlap by an amount x , such that $s = D - x$. This matrix-supported geometry limits applicability of the model to quartz contents of 39–79%, or perhaps 30–85% if the clasts are not perfectly cylindrical. On average, the phyllosilicates are aligned parallel to the (Y) shear band boundaries (i.e., parallel to the macroscopic shear stress τ_m and shear strain rate $\dot{\gamma}_m$), but locally anastomose around the clasts. Given the assumed vertical overlap of adjacent clast layers, i.e., sufficient quartz content, a throughgoing planar foliation is absent.

Within the model microstructure, shear deformation can occur either within the “clast body” zones containing a horizontal phyllosilicate foliation (Type B zones, Figure 3a) or in the “clast overlap” regions containing anastomosing phyllosilicates (Type O zones, Figure 3a). The mechanisms envisaged to be active in these zones, and the way in which they are coupled, are shown in the mechanical analogue diagram in Figure 4. Slip on the phyllosilicate foliation is assumed to be purely frictional (i.e., rate independent) and represented by sliders (elements 1a, 2a, and 3). Thermally activated, rate-dependent shear deformation of the quartz clast bodies and overlaps, for example by SCC or pressure solution, is represented by dashpots (elements 1b and 2b). The horizontal foliation in the Type B zones abuts against the quartz clasts, so that sliding on this foliation (element 1a in Figure 4) requires shear of the clast “bodies” (element 1b). By contrast, in the Type O zones, the foliation anastomoses around the clast “overlaps”. In these zones, deformation can occur either by slip on the phyllosilicates at the zone margins (element 2a) accommodated by shearing of the clast overlaps (element 2b), or by slip on the curved foliation accompanied by dilatation at extensional clast-matrix interface sites (Figure 3a and element 3 in

Figure 4). Since sliding on the foliation is assumed to be a purely frictional process, slip on the curved foliation will not occur unless a critical value of the macroscopic shear stress, τ_{dil} , is attained. When slip is activated, it will cause dilatation and porosity development. We assume that developing porosity concentrates at the extensional quartz-illite interfaces (Figure 3, cf. Figure 2), resulting in a decrease in the clast overlap distance x and hence in the mean inclination of the curved foliation. This in turn causes a decrease in the rate of dilation per unit horizontal displacement on the inclined foliation, or, in the terminology of soil mechanics [Bolton, 1986], a decrease in the dilatation angle Ψ_{dil} with increasing porosity. Following Niemeijer and Spiers [2007], we assume that the appearance of porosity, via clast/matrix debonding, initiates compaction by thermally activated deformation of the clasts, which accelerates as porosity increases. At steady state, dilation due to slip on the curved foliation and compaction by the thermally activated mechanism must balance. This competition between dilatation and compaction is of key importance since it will lead to higher steady state porosities, a flatter foliation, and lower frictional strength as sliding velocity increases, and hence to velocity-weakening slip. Dilatation will continue until a limiting or critical state porosity is reached [cf. Niemeijer and Spiers, 2007].

2.3. Microstructural State Variables

With reference to the assumed steady state microstructure represented in Figure 3b, we now define some additional microstructural descriptors. Consider first the curved foliation. Shear on this will result in an increase of L_2 at the expense of L_1 , as pores open at the extensional clast interfaces. However, since steady state porosities are expected to be small under megathrust conditions, we take $L_1 \approx L_2 \approx L/2$. From Figure 3b, it is now easily shown that

$$L = \frac{k_f \pi D^2}{(D - x_0) f_{qtz}} \quad (1)$$

where k_f is a factor accounting for clast shape (1/4 for cylindrical clasts), x_0 is the vertical overlap of the clasts at zero porosity, and f_{qtz} is the volume fraction of quartz clasts. The vertical clast overlap at zero porosity can hence be written as

$$x_0 = D \left(1 - \sqrt{\frac{k_f \pi}{2 f_{qtz}}} \right) \quad (2)$$

As porosity ϕ increases due to dilational slip on the curved foliation, this overlap decreases from x_0 to an instantaneous value x according to the relation $x = (x_0 - \phi D)/(1 - \phi)$. The decrease in overlap in turn leads to a decrease in the width (λ) of overlapping clast segments (Figure 3b), given $\lambda = 2\sqrt{Dx - x^2}$. In addition, the amplitude and hence inclination of the curved foliation decrease, producing a decrease in the dilatation angle Ψ_{dil} represented in Figure 3b. We treat the curved foliation as a sine function with period L and amplitude $x/2$, on which slip-induced dilatation is controlled by the steepest portion located at the horizontal position $A_x = 0$. The dilatation angle Ψ_{dil} is then given by

$$\tan \Psi_{dil} = \frac{\partial[(x/2) \sin(2\pi A_x/L)]}{\partial A_x} \bigg|_{A_x=0} = \sqrt{\frac{\pi f_{qtz}}{2 k_f}} - \frac{\pi}{2(1 - \phi)} = \frac{\partial \epsilon_{dil}}{\partial \gamma_{dil}} \quad (3)$$

where ϵ_{dil} is the vertical dilational strain and γ_{dil} is the horizontal shear strain produced by dilational slip on the foliation. This angle (Ψ_{dil}) decreases with increasing porosity, reaching zero at a limiting or “critical state” porosity [e.g., Gerogiannopoulos and Brown, 1978; Rutter and Glover, 2012], defined $\phi_c = x_0/D$ when $x = 0$.

The above implies that the model microstructure is characterized by 13 state variables in total, namely, D , s , L , L_1 , L_2 , x , x_0 , λ , f_{qtz} , ϕ , ϕ_c , Ψ_{dil} , and k_f . Of these, only three are independent, e.g., ϕ , D , and f_{qtz} . Note that Ψ_{dil} is a key microstructural state variable that is dynamically determined, through the model, by the imposed deformation conditions and sliding velocity.

3. Macroscopic Shear Stress Balance and Kinematic Relations

These are derived using the mechanical analogue diagram in Figure 4.

3.1. Non-dilatant Deformation

At sufficiently low sliding velocities or at high temperatures, thermally activated shear deformation (creep) of the quartz clasts will be easy. The total resistance to slip on the horizontal foliation will then be lower than the shear stress τ_{dil} needed to activate slip and dilatation on the anastomosing foliation. Under these conditions,

non-dilatant deformation takes place by the parallel processes of (1) slip on the horizontal foliation with serial shear of the clast bodies in the B zones of the microstructure (elements 1a and 1b in Figure 4) plus (2) slip on the horizontal phyllosilicates with serial shear of clast overlaps at the margin of the overlap (O) zones (elements 2a and 2b in Figure 4). Equilibrium between the shear stresses supported by the B and O zones requires $\tau_m = \tau_B = \tau_O$. In addition, the parallel shear processes (1 and 2) operating in the O and B zones mean that the total, measured shear strain rate during non-dilatant deformation of a representative volume of gouge (integer number of unit cells) is given $\dot{\gamma}_m = \dot{\gamma}_B + \dot{\gamma}_O$, where $\dot{\gamma}_B$ and $\dot{\gamma}_O$ denote the shear strain rate contributed to the unit cell by each zone, respectively (i.e., $\dot{\gamma}_B$ and $\dot{\gamma}_O$ are determined by taking into account the thickness of the B or O zone relative to the unit cell thickness). The serial coupling of rate-independent slip on the phyllosilicates (elements 1a and 2a) with thermally activated deformation of clasts (elements 1b and 2b) now implies that $\dot{\gamma}_B = \dot{\gamma}_{qtz-b}$ and $\dot{\gamma}_O = \dot{\gamma}_{qtz-o}$ so that $\dot{\gamma}_m = \dot{\gamma}_{qtz-b} + \dot{\gamma}_{qtz-o}$. Here $\dot{\gamma}_{qtz-b}$ and $\dot{\gamma}_{qtz-o}$ are the shear strain rate contributions to the unit cell due to thermally activated deformation of the clast bodies and clast overlaps, respectively. Lastly, since compaction by thermally activated processes is impossible during non-dilatant deformation of the dense microstructure ($\phi = 0$), we can write $\dot{\epsilon}_{comp} = 0$ for the (positive) compaction strain rate occurring normal to the shear plane.

3.2. Dilatant Deformation

At high slip rates or low temperatures, thermally activated shear deformation of the quartz clasts (elements 1b and 2b in Figure 4) will become difficult, leading to an increase in the total resistance to shear on the horizontal foliation, ultimately activating slip on the curved phyllosilicates in the overlap (O) zones of the microstructure (element 3 in Figure 4). Under these conditions, all elements in Figure 4 are active. The measured shear strength is now equal to that required to activate slip on the anastomosing foliation (τ_{dil}), which, coupled with stress equilibrium between B and O zones, means $\tau_m = \tau_{dil} = \tau_B = \tau_O$. The total shear strain rate $\dot{\gamma}_m$, in turn, is given $\dot{\gamma}_m = \dot{\gamma}_B + \dot{\gamma}_O + \dot{\gamma}_{dil}$, or equivalently $\dot{\gamma}_m = \dot{\gamma}_{qtz-b} + \dot{\gamma}_{qtz-o} + \dot{\gamma}_{dil}$, where $\dot{\gamma}_{dil}$ is the shear strain rate contribution to the unit cell by dilatant slip on the curved phyllosilicates. This mechanism produces an associated dilatational strain rate, $\dot{\epsilon}_{dil}$, which, following the classical soil mechanics approach to granular flow [Bolton, 1986], is written

$$\dot{\epsilon}_{dil} = \left(\frac{d\epsilon_{dil}}{d\gamma_{dil}} \right) \frac{d\gamma_{dil}}{dt} = (\tan \Psi_{dil}) \dot{\gamma}_{dil} \quad (4)$$

The porosity generated will induce compaction by thermally activated deformation of the quartz clasts at a rate $\dot{\epsilon}_{comp}$. Taking compaction as positive, the total, measured compaction strain rate is therefore given $\dot{\epsilon}_m = \dot{\epsilon}_{comp} - \dot{\epsilon}_{dil}$. At steady state, dilatation and compaction must balance, resulting in a steady state porosity corresponding to the condition that $\dot{\epsilon}_m = 0$ or $\dot{\epsilon}_{comp} = \dot{\epsilon}_{dil}$.

4. Microscopic Stress Balance and Shear Resistance to Deformation

4.1. Clast Body Zones

During non-dilatant deformation at low slip rates and/or high temperatures, but also during dilatant deformation, the total shear stress, acting over the shear plane of area LD per unit cell (Figure 3b), is supported, in the B zone of the cell, by slip on the horizontal phyllosilicates plus thermally activated deformation of the clast bodies (element 1, Figure 4), each contributing in proportion to the horizontal area occupied. This means that

$$\tau_B = \tau_{ph} \left(1 - \frac{A_{qtz-b}}{LD} \right) + \tau_{qtz-b} \frac{A_{qtz-b}}{LD} \quad (5)$$

where τ_{ph} is the shear stress needed to drive frictional slip on the horizontal phyllosilicate foliation and τ_{qtz-b} is that needed to drive thermally activated clast body deformation. A_{qtz-b} represents the average horizontal area occupied by a single clast body within zone B of the unit cell and is given $A_{qtz-b} = [(1/4\pi D^2 - 2A'_{seg})D]/(D-2x)$, where $A'_{seg} = [16x^2(D-x) + 3x^3]/[12\sqrt{(Dx-x^2)}]$ [Harris and Stocker, 1998] is the area of an individual clast segment located in the overlap zone of the cell in the plane of Figure 3b.

4.2. Clast Overlap Zones

Similarly, during both non-dilatant and dilatant deformation, the shear stress supported in the overlap (O) region of the unit cell is given by:

$$\tau_O = \tau_{ph} \left(1 - \frac{A_{qtz-o}}{LD} \right) + \tau_{qtz-o} \frac{A_{qtz-o}}{LD} \quad (6)$$

where τ_{qtz-o} is now the shear stress needed to drive thermally activated deformation of the quartz overlaps and $A_{qtz-o} = \lambda D = 2D\sqrt{(Dx-x^2)}$ is the area over which the overlap is displaced by slip at its base.

To quantify τ_{ph} in equations (5) and (6), sliding on the horizontal foliation can be described by a rate-independent, cohesionless friction law given $\tau_{ph} = \tilde{\mu}\sigma_n^{eff}$, where $\tilde{\mu}$ is the friction coefficient within the phyllosilicates and σ_n^{eff} is the macroscopic, effective normal stress. The resistance to shear offered by the quartz clast bodies and overlaps (τ_{qtz-b} and τ_{qtz-o}) depends on the mechanism by which thermally activated deformation occurs. In principle, this could be any creep mechanism, including SCC and pressure solution, though later, we assume it is pressure solution.

4.3. Stress Needed for Dilatant Deformation

The shear stress needed to activate dilatant slip on the curved foliation in the overlap zones (O) can be obtained from Figure 3b by considering the balance of forces acting on the anastomosing foliation. We approximate the mean trajectory of the curved foliation by straight lines (ℓ_1 and ℓ_2) joining the tops and bottoms of adjacent clasts, as shown in Figure 3b. For the mean inclination (Ψ_{fr}) of the curved foliation, this yields

$$\tan \Psi_{fr} = \frac{2(D - x_0)f_{qtz}}{k_f \pi D^2} x \quad (7)$$

which, since $L_1 \approx L_2$, holds for both halves of the unit cell. Balancing the normal and shear forces on ℓ_1 now yields

$$S_{n1}\ell_1 D = \sigma_n^{eff} L_1 D \cos \Psi_{fr} + \tau_1 L_1 D \sin \Psi_{fr} \quad (8)$$

$$S_{s1}\ell_1 D = \tau_1 L_1 D \cos \Psi_{fr} - \sigma_n^{eff} L_1 D \sin \Psi_{fr} \quad (9)$$

where S_{n1} and S_{s1} are the normal and shear stress components on ℓ_1 , and τ_1 is the portion of the macroscopic shear stress that acts over distance L_1 . The normal and shear stress components on ℓ_1 are linked via $S_{s1} = \tilde{\mu} S_{n1}$. Using $\cos \Psi_{fr} = L_1/\ell_1$, equations (8) and (9) thus yield $\tau_1 = [(\tilde{\mu} + \tan \Psi_{fr})/(1 - \tilde{\mu} \tan \Psi_{fr})]\sigma_n^{eff}$.

The portion of the shear stress τ_2 acting over distance L_2 can be obtained similarly. The macroscopic shear strength is subsequently obtained by adding τ_1 and τ_2 , accounting for the lengths over which they act. This yields

$$\tau_{dil} = \left\{ \frac{\tilde{\mu}(1 + \tan^2 \Psi_{fr})}{1 - \tilde{\mu}^2 \tan^2 \Psi_{fr}} \right\} \sigma_n^{eff} \quad (10)$$

for the shear stress required for slip on the curved foliation.

5. Thermally Activated Deformation of Quartz

As already explained, SCC is too poorly constrained to model clast deformation by this mechanism. Data on the kinetics of SCC are available only up to 300°C in the presence of water vapor at sub-atmospheric pressures and show extreme sensitivity to experimental conditions [Atkinson, 1984]. We therefore focus on pressure solution as the thermally activated mechanism operating in our model, assuming that it is controlled by the interfacial reactions of dissolution and precipitation. A simple order of magnitude calculation shows that for fine grained materials such as gouges, diffusion is too fast to be rate controlling. The required flow laws for clast shearing and compaction (pore filling) by pressure solution have been derived by Den Hartog [2013], assuming that the two processes can be treated independently [cf. Niemeijer and Spiers, 2007; Raj, 1982]. Consistent with Figure 4, Den Hartog [2013] further assumed homogeneous simple shear of the clast bodies (B zones) and a fixed shape of the overlapping portions of the clasts (O zones). In addition, pressure solution was taken to be the sole dissipative process associated with shear of the clasts, allowing the dissipation due to the rate controlling steps of dissolution and precipitation to be equated to the mechanical work of shearing [cf. Bos and Spiers, 2002; Lehner and Bataille, 1984]. Beyond this, standard treatments of pressure solution were followed [Raj, 1982; Shimizu, 1995; Spiers et al., 2004; see also Bos and Spiers, 2002]. The results for clast body shear, clast overlap shear, and compaction by pressure solution are, respectively

$$\dot{\gamma}_{qtz-b} = \frac{A l \tau_{qtz-b} \Omega}{RT} \frac{D - 2x}{D(D - x)} \quad (11)$$

$$\dot{\gamma}_{qtz-o} = \frac{2 l \tau_{qtz-o} \Omega}{RT} \frac{1}{\sqrt{Dx - x^2}} \quad (12)$$

$$\dot{\epsilon}_{comp} = \frac{2 l \sigma_n^{eff} \Omega}{RT} \frac{A_{pore}}{(D - x)DL} \quad (13)$$

where A is a shape factor that takes values between 2 and π , l is the product of the dissolution rate coefficient k_+ and molar volume Ω of quartz, and R is the gas constant. All the above flow laws apply to a unit cell and are scaled to account for the thickness of the zone in which the process is active relative to the unit cell thickness. Compaction occurs by pressure solution transfer from compressively stressed illite-quartz interfaces to debonded (dilated) interfaces (pore walls) with surface area A_{pore} . This area is described using a power law function of ϕ , written $A_{pore} = (A_{pore-c}/2)(\phi/\phi_c)^n$, where ϕ_c and A_{pore-c} are the porosity and pore area per clast at the critical state. The exponent $n < 1$ is used to account for a large pore length to width ratio at the onset of dilatation (clast interface debonding), evolving to a smaller ratio with increasing steady state porosity, consistent with the observed microstructure (Figure 2). Note that $A_{pore-c} = (\pi D^2)/2$, which is the maximum geometrically possible debonded pore area, i.e., one half of the cylindrical clast surface.

6. Final Model Assembly and Implementation

We now combine the foregoing equations to provide a framework for calculating the macroscopic, steady state friction coefficient ($\mu = \tau_m/\sigma_n^{eff}$) as a function of shear strain rate or slip velocity.

6.1. Determining Shear Strength as a Function of Slip Velocity

6.1.1. Non-dilatant Deformation

During non-dilatant deformation at low slip velocities and/or high temperatures, the total shear stress supported is defined as $\tau_m = \tau_B = \tau_O$ and the shear strain rate as $\dot{\gamma}_m = \dot{\gamma}_B + \dot{\gamma}_O$ (elements 1 and 2, Figure 4). To obtain τ_m as a function of $\dot{\gamma}_m$, we impose $\dot{\gamma}_m$, express $\dot{\gamma}_O$ in terms of $\dot{\gamma}_m$ and $\dot{\gamma}_B$, and calculate $\dot{\gamma}_B = \dot{\gamma}_{qtz-b} = \dot{\gamma}_m - \dot{\gamma}_O$. We subsequently use $\dot{\gamma}_{qtz-b}$ to determine τ_{qtz-b} via the flow law given in equation (11). The value of τ_{qtz-b} obtained, combined with $\tau_{ph} = \tilde{\mu}\sigma_n^{eff}$, then yields $\tau_B = \tau_m$ through equation (5). By incrementing $\dot{\gamma}_m$, we can then thus calculate corresponding values of $\dot{\gamma}_{qtz-b}$ and $\tau_B = \tau_m$, yielding the desired description of the total shear strength as a function of the imposed shear strain rate.

6.1.2. Dilatant Deformation

During dilatant deformation at high slip rates or low temperatures, $\tau_m = \tau_{dil} = \tau_B = \tau_O$ and $\dot{\gamma}_m = \dot{\gamma}_B + \dot{\gamma}_O + \dot{\gamma}_{dil}$ (refer Figure 4). To calculate τ_m as a function of $\dot{\gamma}_m$, we increment the porosity from 0 to ϕ_c and calculate the corresponding values of $\tan\Psi_{dil}$ and $\tan\Psi_{fr}$ using equations (3) and (7). Using $\tan\Psi_{fr}$, equation (10) gives τ_{dil} . The corresponding shear strain rate $\dot{\gamma}_m$ is calculated via $\dot{\gamma}_m = \dot{\gamma}_B + \dot{\gamma}_O + \dot{\gamma}_{dil}$, realizing that $\dot{\gamma}_B = \dot{\gamma}_{qtz-b}$ and $\dot{\gamma}_O = \dot{\gamma}_{qtz-o}$ and using the flow laws in equations (11) and (12). Here $\dot{\gamma}_{dil}$ is obtained via equation (4) and using the steady state condition $\dot{\epsilon}_{comp} = \dot{\epsilon}_{dil}$, where $\dot{\epsilon}_{comp}$ is calculated using equation (13).

6.2. Model Input Data

To obtain a clear picture of the model characteristics, while allowing for comparison with our experimental results for illite-quartz gouge (Figure 1) [see also Den Hartog and Spiers, 2013], we applied the model to predict the steady state frictional behavior of illite-quartz gouge for typical experimental conditions, i.e., an effective normal stress of 170 MPa, temperatures of 100–600°C, and sliding velocities of 1–100 $\mu\text{m/s}$. Besides these variables, our model requires specification of the microstructural state variables D and f_{qtz} , the pore shape evolution parameter n , the phyllosilicate friction coefficient $\tilde{\mu}$, and the pressure solution rate coefficient l .

For the present calculations, a grain size D of 0.93 μm was used, being the average grain size observed in the illite-quartz samples after deformation [Den Hartog et al., 2012a]. Sliding velocity (V) and macroscopic shear strain rate $\dot{\gamma}_m = V/w$ were connected using an average sample thickness (w) of 0.57 mm. This is the average sample thickness for all experiments conducted at $\sigma_n^{eff} = 170$ MPa. Individual sample thicknesses were obtained from the final thickness and axial compression measured during the experiments. We assume that shear deformation was uniformly distributed throughout the samples. The average quartz fraction (f_{qtz}) during the experiments was estimated from the initial mass fraction of 35% (≈ 35 vol% for quartz-illite), assuming that thinning during shear was caused by preferential loss of illite [cf. Den Hartog et al., 2012a]. This yielded an average quartz fraction of 56% for all experiments conducted at 170 MPa effective normal stress. A mean clast overlap at zero porosity of 0.15 μm was obtained from the grain size (D) and putting $f_{qtz} = 56\%$ (equation (2)). This in turn yielded an average clast spacing L of 1.55 μm (equation (1)) and a critical state porosity ($\phi_c = x_o/D$) of 16%. Based on microstructural observations, we chose the exponent n in equation (13) to be 0.3. This choice leads to a dependence of pore wall area on porosity consistent with a decrease in pore aspect ratio from $\sim 50:1$ upon initial formation (i.e., upon clast-matrix debonding) to $\sim 2.5:1$ at the critical state porosity.

Figure 5. Evolution of the steady state friction coefficient μ with sliding velocity V predicted by the current model for different values of (a) temperature, (b) effective normal stress, (c) quartz fraction, and (d) quartz clast grain size, at otherwise fixed conditions. The values of the quantities held constant are specified in the top left corner of each panel. The regions of non-dilatant versus dilatant behavior are indicated for the curve at 300°C in Figure 5a.

To describe the kinetics of pressure solution, i.e., to obtain $I = k_+ \Omega$, we used the empirical equation for the dissolution rate coefficient (k_+) for quartz, based on geometric surface area, provided by *Tester et al.* [1994]. This relation represents a compilation of data on k_+ versus T and was chosen because pressure solution rates in our model are determined by the geometric surface area available for dissolution and precipitation at clast surfaces (as opposed to Brunauer, Emmett, and Teller or BET surface area). To obtain $\tilde{\mu}$ for illite as a function of temperature, we used the effect of temperature on friction coefficient observed by *Den Hartog et al.* [2013] for its close structural analogue pure muscovite, plus the friction coefficient of 0.3 for pure illite measured at room temperature by *Tembe et al.* [2010]. This yielded the relation $\tilde{\mu} = 0.303 + (T - 20) \cdot 5.375 \cdot 10^{-4}$.

7. Model Predictions and Comparison With Experimental Results

In the following, we illustrate the principal trends predicted by our microphysical model, and we compare the results with the trends in steady state strength and in the RSF parameter ($a-b$) previously observed in experiments on illite-quartz gouge by *Den Hartog and Spiers* [2013] (see also Figure 1).

7.1. Dependence of Shear Strength (μ) on Deformation Conditions and Microstructural Variables

7.1.1. Effect of Sliding Velocity V

The dependence of gouge friction coefficient μ on sliding velocity V predicted by the model for the chosen reference conditions at steady state is shown in Figure 5a. At low sliding velocities, non-dilatant deformation controls friction, with slip on the horizontal foliation being accommodated by pressure solution of the quartz clasts. At the lowest velocities explored, μ approaches an asymptotic value, determined by the shear strength $\tilde{\mu}$ of the horizontal phyllosilicates (taking into account the relevant area fraction), since easy pressure solution of the clasts under these conditions contributes negligibly to the strength. With increasing velocity, pressure solution becomes more difficult, resulting in an increase in steady state friction coefficient μ (i.e., velocity-strengthening behavior). This increase in μ continues until slip is activated on the anastomosing foliation

Figure 6. Trends predicted by the current model for the dependence of steady state friction coefficient μ on (a) temperature, (b) effective normal stress, (c) quartz fraction, and (d) quartz grain size, under otherwise constant conditions. The values of the quantities held constant are specified in the top left corner of each panel. Discontinuities related to the switch from dilatant to non-dilatant behavior with increasing temperature in the model are indicated where relevant, with a dotted line in the same gray shade as the corresponding curve, with the regions corresponding to non-dilatant (nd) and dilatant (d) behavior indicated.

(at the peak of the μ versus V curve). Dilatant deformation (i.e., porosity development) ensues and in turn activates compaction by pressure solution. Slip-induced dilatation and pressure solution compaction now compete with each other, balancing at steady state. The steady state porosity is of course higher at higher velocities, resulting in a lower Ψ_{dil} value and a lower shear strength μ (equations (7) and (10)), hence velocity-weakening behavior (Figure 5a). At the highest velocities, the porosity approaches the critical state value and the friction coefficient approaches $\tilde{\mu}$, as the phyllosilicates align horizontally.

Note that the present model predicts a similar μ versus V profile to that predicted by the Bos-Niemeijer-Spiers model [2001, 2005, 2007] for a clast-supported microstructure. There are two key differences, though. First, the magnitude of velocity weakening is much lower here. Second, the present model predicts the transition from non-dilatant to dilatant deformation, whereas in the Bos-Niemeijer-Spiers model this was assumed to be achieved via an independent dilatation criterion resembling Byerlee's rule.

Our previous experiments were conducted at only three velocities (1, 10, and 100 $\mu\text{m/s}$). We accordingly focus henceforth on comparing the predicted versus measured effects of T , σ_n^{eff} , and quartz fraction (f_{qtz}) on μ and on μ versus V behavior. Figures 5a–5d show how the μ versus V curve shifts with T , σ_n^{eff} , f_{qtz} , and D . The resulting trends in μ versus T , σ_n^{eff} , f_{qtz} , and D at constant V are shown in Figures 6a–6d. These trends are described below and compared with our experimental observations.

7.1.2. Effect of T

The predicted effect of temperature on steady state friction coefficient is shown in Figure 6a, at sliding velocities of 1, 10, and 100 $\mu\text{m/s}$. The main trend visible is an increase in μ with T , which is in good agreement with the μ versus temperature trend observed in our experiments (Figure 1b). However, at 1 and 10 $\mu\text{m/s}$, μ is predicted to decrease sharply at the highest temperatures. This was not observed in our experiments, perhaps indicating that the discontinuous transition between the non-dilatant and dilatant regimes in our

model (Figure 5) is smoothed out in real materials and in experiments performed at discrete intervals in T . In addition, the limited microstructural evidence for pressure solution suggests that the full transition from dilatant to non-dilatant behavior did not occur in our experiments. The possibility of a smoothed out transition is further supported by the experimental data for a shear strain of 50 (Figure 1b) [see also *Den Hartog et al.*, 2012a], which do show some flattening off in μ versus T above $\sim 350^\circ\text{C}$.

7.1.3. Effect of σ_n^{eff}

The dependence of μ on σ_n^{eff} at a fixed sliding velocity of $10\ \mu\text{m/s}$ is shown in Figure 6b at temperatures of 450, 500, and 550°C . This shows that σ_n^{eff} has little effect on μ below 500°C in broad agreement with the experimental observations (Figure 6b versus Figure 1b) except at the lowest temperatures and normal stresses. Note, however, that the decrease in μ with increasing σ_n^{eff} observed under these conditions (Figure 1b) was attributed by *Den Hartog and Spiers* [2013] to sample cohesion at low effective normal stresses—an effect not incorporated in the model. At 550°C , μ is predicted to decrease with σ_n^{eff} but does not. This may reflect the fact that at 550°C , the predicted effect of σ_n^{eff} on μ is evaluated at a fixed sliding velocity that falls near the maximum in the μ versus V curve (cf. Figure 5a), where the discontinuous transition from non-dilatant to dilatant behavior results in sharp changes in μ with σ_n^{eff} . Since the transition is likely smoothed out in real materials, discrepancies between model and experiment are unsurprising in this region.

7.1.4. Effect of f_{qtz}

The dependence of the friction coefficient on quartz fraction at $10\ \mu\text{m/s}$ and at 450, 500, or 550°C is shown in Figure 6c. Depending on temperature, an increase in quartz fraction either results in no changes in μ (450°C and below), a decrease in μ at low quartz contents followed by a slight increase (500°C), or an increase in μ (550°C and above). These different responses are related to the position on the μ versus V curve, i.e., on the selected slip velocity in combination with temperature (refer Figures 5a and 5c). The experimentally observed increase in μ with f_{qtz} seen by *Den Hartog and Spiers* [2013] at 140°C (the only temperature tested, see Figure 1b) and by others at room temperature [*Brown et al.*, 2003; *Tembe et al.*, 2010] is clearly not reproduced by the model at low temperatures ($\leq 450^\circ\text{C}$). This failure in the predictions (i.e., in Regime 1, Figure 1b) is related to the approach to critical state behavior in which the model predicts zero overlap of quartz clasts, hence no effect of f_{qtz} . The fact that experiments do show a positive effect of f_{qtz} on frictional strength suggests that in reality critical state behavior may not require zero overlap of quartz clasts ($x=0$) but involves a rate-independent balance between pore opening and closure due to slip on the anastomosing phyllosilicate foliation at a lower value of ϕ_c than used in the model. No experimental data are available for comparison with the increase in μ with f_{qtz} predicted by the model at 500°C and above (Figure 6c).

7.1.5. Effect of D

Although not addressed in our experiments on illite-quartz gouge, we also investigated the effect of quartz grain size on the steady state μ versus velocity curve predicted by the model—see Figure 6d. At intermediate velocities, an increase in D leads to an increase in μ at small grain sizes, followed by a slight decrease toward a constant friction level at large grain sizes, reflecting the transition from non-dilatant to dilatant behavior (Figure 6d).

7.2. Trends in the RSF Parameter ($a-b$)

The velocity dependence of friction observed in experiments is usually quantified in the framework of the rate and state dependent friction model (RSF) [*Dieterich*, 1978, 1979; *Ruina*, 1983], using the parameter

$$(a-b) = \frac{\Delta\mu_{ss}}{\Delta \ln V} = \frac{\partial\mu_{ss}}{\partial \ln V} \quad \text{in the limit } \Delta \ln V \rightarrow 0 \quad (14)$$

[e.g., *Marone*, 1998; *Scholz*, 1998]. Here a represents the absolute magnitude of the instantaneous change in apparent friction coefficient μ upon a step change in sliding velocity from a reference value V_0 , characterized by a steady state friction coefficient μ_{ss} , to a new velocity V , while b reflects the magnitude of the change in μ during evolution to a new steady state. The overall change in steady state friction coefficient is denoted $\Delta\mu_{ss}$. Positive ($a-b$) values indicate an increase in friction coefficient with increasing velocity, i.e., velocity strengthening, which ensures stable slip behavior, while negative ($a-b$) values signify velocity-weakening and potentially unstable behavior [e.g., *Scholz*, 1998]. Almost all previous experiments on illite-quartz gouge have made use of ($a-b$) to investigate the velocity dependence of frictional behavior [*Ikari et al.*, 2009; *Morrow et al.*, 1992; *Saffer and Marone*, 2003; *Tembe et al.*, 2010], determining ($a-b$) by stepping the sliding velocity, usually in order of magnitude jumps (e.g., 1, 10, and $100\ \mu\text{m/s}$ [*Den Hartog et al.*, 2012a]), or else half-order jumps [*Ikari et al.*, 2009].

Figure 7. Plots of $(a-b)$ versus T predicted by the current model, showing dependence on (a) slip velocity, (b) effective normal stress, (c) quartz fraction, and (d) grain size, under otherwise constant conditions. The values of the quantities held constant are specified in the top left corner of each panel. All curves are derivative based, except the two bold curves in Figure 7a, which we obtained by taking the finite difference between the strengths predicted at 1 and 10 $\mu\text{m/s}$ or at 10 and 100 $\mu\text{m/s}$. The discontinuity related to the switch from dilatant to non-dilatant behavior with increasing temperature in the model, visible in the derivative-based $(a-b)$ curves, is indicated with a dotted line for each curve, in the same gray shade as the corresponding curve, with the regions corresponding to non-dilatant (nd) and dilatant (d) behavior indicated. In real materials, or in experiments where $(a-b)$ is determined from finite (10X) velocity steps, such a discontinuity is expected to be smoothed out.

7.2.1. Effects of V and T

We determined $(a-b)$ from the predicted μ versus V curves (Figure 5a) by direct application of equation (14), at fixed velocity, and fixed temperatures in the range 100–600°C, to obtain a finite difference approximation of the continuous derivative of $\mu(V)$, i.e., of $\partial\mu/\partial V$, taking $\mu = \mu_{ss}$. The $(a-b)$ values thus calculated at velocities of 1, 10, and 100 $\mu\text{m/s}$ are shown versus temperature in Figure 7a. Note the jump in $(a-b)$ at the discontinuous transition between dilatant and non-dilatant behavior predicted by the model (Figure 7a). In reality, such a discontinuity will be smoothed out by factors such as distributed grain size and distributed dilatation angle. Moreover, in experiments, it will generally be smoothed out by the use of velocity steps to determine $(a-b)$. Three regions of behavior can be distinguished in Figure 7a: velocity-neutral behavior at low temperatures, velocity weakening at intermediate temperatures, and velocity-strengthening behavior at high temperatures.

We also determined $(a-b)$ versus T as predicted by the model for jumps in V from either 1 to 10 $\mu\text{m/s}$ or 10 to 100 $\mu\text{m/s}$ (Figure 7a), thus simulating the velocity stepping approach to determining $(a-b)$ used in our previous experiments on illite-quartz gouge [Den Hartog and Spiers, 2013; Den Hartog et al., 2012a]. These profiles show the same three regimes of $(a-b)$ versus temperature, but with reduced velocity weakening, a shift of the velocity-weakening regime to slightly higher temperatures (+25°C) and smoothing out of the discontinuity. Note that all predicted $(a-b)$ curves shown in Figure 7a show a clear dependence on slip velocity, shifting toward higher temperatures with an increase in slip velocity. In addition, velocity weakening is more pronounced at higher slip velocity, i.e., $(a-b)$ values are more negative.

The predicted existence of three regimes of $(a-b)$ with increasing temperature is in good agreement with our experimental observations on illite-quartz mixtures (Figure 7a versus Regimes 1, 2, and 3 in Figure 1a).

The magnitude of $(a-b)$, i.e., $|a-b|$, shows reasonable agreement too, in the velocity-weakening regime, i.e., in Regime 2. The predicted displacement of the $(a-b)$ versus T profile toward higher temperatures with increasing sliding velocity also agrees with the trends observed in the experiments (Figures 1a, 1c, and 1e).

However, the temperature range of velocity-weakening behavior in Regime 2 in our experiments was 250–400°C (Figure 1a). This range overlaps with those predicted in Figure 7a for finite velocity steps involving 1 and 10 $\mu\text{m/s}$, but extends to lower temperatures. A possible explanation for this discrepancy lies in our assumption of homogeneous shear in the fault gouge model. If, in the experiments, some degree of localization occurs, then the predicted μ versus V curves should be displaced to higher V for proper comparison (cf. Figure 5), while the predicted $(a-b)$ versus T profiles should be shifted to lower temperatures. Note that a factor 10X increase in strain rate produced by shear localization would shift the velocity-weakening regime down in temperature by $\sim 100^\circ\text{C}$ in the velocity range of 1–100 $\mu\text{m/s}$ (Figure 7a).

A further discrepancy between model predictions and experiments concerns the velocity dependence of μ , i.e., $(a-b)$, at low temperatures (i.e., in Regime 1). Although similar three-regime $(a-b)$ versus temperature behavior is seen in our experimental data (Figure 1) and in the model predictions (Figures 1a and 7a), $(a-b)$ values at the lowest temperatures are positive in the experiments but zero in the predictions. This may be related to the velocity-strengthening behavior typically reported for pure phyllosilicates sheared wet at low temperatures [e.g., *Ikari et al.*, 2011] and recently observed by *Den Hartog et al.* [2013] for pure muscovite, notably at 200 and 600°C (at 400°C near neutral behavior was observed). The present model assumes that $\tilde{\mu}$ for pure illite is velocity independent. Incorporating a velocity-strengthening $\tilde{\mu}$ would result in the observed velocity-strengthening behavior at low temperatures. In addition, if pure illite shows a near neutral minimum in $(a-b)$ at intermediate temperatures, similar to pure muscovite, incorporation of such an effect into the model would tend to produce a broadening of the predicted temperature range of velocity weakening, more closely approaching the experimental observations.

In Regime 3 (velocity strengthening at high T), both model predictions and experimental observations show positive $(a-b)$. The predicted magnitude of $(a-b)$ for the derivative-based curves (Figure 7a), however, is much larger than that observed in the experiments (Figure 1a). Comparison with the curves in which $(a-b)$ is obtained using the finite jumps in velocity (Figure 7a), used in the experimental determinations, is more relevant, but still shows predicted values $\sim 5\text{X}$ larger than observed experimentally (Figure 7a versus Figure 1a). Once again, this could reflect the discontinuous nature of the transition from non-dilatant to dilatant deformation in the model, which in reality is expected to be smoothed out.

7.2.2. Effects of σ_n^{eff} , f_{qtz} , and D

The predicted effect of σ_n^{eff} on the $(a-b)$ versus temperature profile is shown in Figure 7b, using the derivative of the corresponding μ versus V curve taken at 10 $\mu\text{m/s}$. With increasing effective normal stress, the three temperature-dependent $(a-b)$ regimes systematically shift toward lower temperatures. This qualitative trend is in good agreement with our experimental observations on illite-quartz mixtures (cf. Figure 1d).

The predicted dependence of the $(a-b)$ versus T curve on quartz content is shown in Figure 7c, again using the derivative of the relevant μ versus V curves determined at 10 $\mu\text{m/s}$. In line with the effects of increasing quartz content observed in Figure 5c, when the quartz fraction increases from 0.56 to 0.7, $(a-b)$ becomes increasingly negative in the velocity-weakening regime, but the negative region still spans the same temperature range. A decrease in quartz content below 56% reduces the velocity-weakening response and, for a quartz fraction of 0.4, completely removes the velocity-weakening region (cf. Figure 5c). In the experiments at 140°C reported by *Den Hartog and Spiers* [2013], an increase in quartz content was observed to lead to a decrease in $(a-b)$, which was interpreted to imply a shift of the $(a-b)$ profile versus T profile toward lower temperatures (Figure 1e). However, the vertical exaggeration of the $(a-b)$ versus temperature profile with increasing quartz content predicted by our model is also in agreement with our experimental results [*Den Hartog and Spiers*, 2013].

Profiles of $(a-b)$ versus temperature predicted for different quartz clast grain sizes are displayed in Figure 7d. With increasing grain size, the velocity-weakening region shifts toward higher temperatures, in line with the displacement of the μ versus V curves toward lower velocities with increasing grain size, hence decreasing pressure solution rate, as predicted in Figure 5d.

Figure 8. Model predictions for the dependence of $(a-b)$ on temperature at slip rates of 10^{-9} , 10^{-8} , and 10^{-7} m/s, expected to be relevant to earthquake nucleation (plate velocity V_{pl} , $10V_{pl}$, and $100V_{pl}$). The discontinuity at the boundary between dilatant and non-dilatant behavior is marked with a dotted line, in the same gray shade as the corresponding $(a-b)$ versus temperature curve (see also Figure 7). The effective normal stress, quartz fraction, and grain size were held constant, as indicated.

8. Implications for Subduction Megathrust Seismogenesis

To predict the depth range of velocity-weakening behavior, i.e., a negative $(a-b)$, under in situ subduction megathrust conditions, depth profiles of all relevant quantities, notably temperature and effective normal stress, are needed. However, although the temperature is reasonably well constrained for various subduction zones [Peacock, 2009; Peacock and Wang, 1999; Spinelli and Wang, 2008], the pore fluid pressure versus depth profile, and thus the effective vertical and hence normal stress versus depth profile, is largely a matter of speculation at present [Den Hartog and Spiers, 2013; Saffer and Tobin, 2011]. Though the effect of σ_n^{eff} on μ is probably relatively small, at least in the velocity-weakening regime (see Figures 1b and 6b), this makes the

construction of useful μ and $(a-b)$ versus depth profiles for natural subduction megathrusts problematic, beyond the simple assumption that σ_n^{eff} has no effect, or that P_f is hydrostatic ($P_f = \bar{\rho}_f g h$) or a fixed fraction of the lithostatic pressure [e.g., $P_f/P_{litho} = 0.9-0.98$; Seno, 2009].

Like the temperature, the slip rate associated with earthquake nucleation is somewhat better constrained, generally being assumed to be within $\sim 1-2$ orders of magnitude of plate velocities of $\sim 10^{-9}$ m/s [e.g., Segall and Rice, 2006] and thus 3–5 orders of magnitude lower than applied in our experiments and model predictions. Den Hartog and Spiers [2013] attempted to extrapolate the experimentally derived $(a-b)$ versus temperature profile to these natural rates, using the observed trends in $(a-b)$ versus V , assuming no influence of departures in σ_n^{eff} from the experimental value of 170 MPa. However, the results were too poorly constrained to be reliable.

The present microphysical model provides an improved basis for extrapolation of the $(a-b)$ versus temperature trend to sub-experimental slip velocities. The predicted dependence of $(a-b)$ on temperature is shown in Figure 8 for slip velocities of 10^{-9} , 10^{-8} , and 10^{-7} m/s, using otherwise the same conditions as in our reference case experiments ($\sigma_n^{eff} = 170$ MPa, $f_{qtz} = 0.56$, $D = 0.93$ μm , sample thickness = 0.57 mm, and $n = 0.3$). These predictions imply velocity-weakening behavior at temperatures of $\sim 140-193^\circ\text{C}$, $190-245^\circ\text{C}$, and $230-310^\circ\text{C}$, at 10^{-9} , 10^{-8} , and 10^{-7} m/s, respectively. The temperature of 140°C for onset of velocity-weakening behavior at 10^{-9} m/s agrees well with the updip seismogenic limit on subduction megathrusts, but, with the transition back to velocity-strengthening occurring at 193°C , the temperature range of velocity weakening is narrower than that of $150-350^\circ\text{C}$ over which seismogenesis occurs on megathrusts [Hyndman et al., 1997]. This may imply that velocities higher than 10^{-9} m/s should be taken into account, in relation to rupture nucleation (cf. Figure 8). However, as we emphasized previously [Den Hartog et al., 2013], illite will be replaced by muscovite at temperatures of $\sim 200-300^\circ\text{C}$ [Hunziker, 1986; Merriman and Frey, 1999; Van de Kamp, 2008], and thus it is the behavior of muscovite-quartz gouge that should determine the temperature at which velocity-weakening behavior changes back to strengthening. Prediction of the temperature associated with this transition requires a microphysical model specifically developed for muscovite-quartz mixtures, which is beyond the scope of the current paper. Here, we note simply that experiments on muscovite-quartz gouge extend the range of velocity weakening seen in illite-quartz gouge upward in temperature by 100°C at similar conditions [Den Hartog et al., 2013].

Following the earlier approach of Den Hartog et al. [2012a, 2012b], based on experimental data only, the temperature range of velocity-weakening behavior at 10^{-9} m/s predicted by the current model can be compared with that assumed in numerical studies, addressing seismogenesis on subduction megathrusts, such as those by Liu and Rice [2005] or Shibazaki and Shimamoto [2007]. As explained above, the transition

from illite to muscovite at ~ 200 – 300°C limits the comparison of the present model for illite-quartz to temperatures below this. In this range, the model predicts the onset of velocity-weakening behavior at $\sim 140^\circ\text{C}$. This is higher than the 100°C assumed by *Liu and Rice* [2005] and *Shibazaki and Shimamoto* [2007] for the southwest Japan subduction zone. In addition, the $(a-b)$ values predicted in the current study have a significantly higher magnitude (i.e., 5X more negative) than those used by *Liu and Rice* [2005] and *Shibazaki and Shimamoto* [2007]. The effects of these higher values should be explored in future modeling work. Further improvements in $(a-b)$ versus depth profiles in the future will require better insight into the pore fluid pressure distribution and hence effective normal stress that may be expected in situ.

Of course, to fully link the current steady state model to classical RSF friction models for modeling megathrust seismogenesis, the present model needs to be generalized to account for transient effects. This can be done by avoiding the steady state assumption of zero net compaction rate, i.e., the assumed balance between dilatation and thermally activated compaction. If, instead, the differential equation describing the net compaction rate as the difference between the dilatation and thermally activated compaction rate is used, then the transient response of the microstructure (ϕ), and hence friction coefficient (μ), to changes in sliding velocity can, in principle, be determined, as well as the classical RSF parameters a , b , and d_c . Moreover, the dependence of a , b , and d_c on V , T , σ_n^{eff} , and quartz content can also be predicted. Obtaining a value for the critical slip distance d_c needed to reach a new steady state friction value after a change in slip velocity would further allow the prediction of the critical stiffness of gouge zones, yielding additional insight into slip stability [e.g., *Beeler et al.*, 2007].

Finally, we note that the good agreement found between the predicted temperature-dependent three-regime behavior and the sequence of aseismic-seismic-aseismic zones seen on natural megathrusts, suggests that seismogenic behavior is caused by the same mechanism that results in velocity-weakening behavior in the model. This would mean that seismogenesis is in fact caused by gouge compaction through thermally activated deformation of quartz clasts, most likely by pressure solution. Without a thermally activated compaction process such as this to compete with rate-independent dilation due to slip on the anastomosing foliation (i.e., due to granular/cataclastic flow), our model implies that velocity weakening would not occur. The implication is that velocity weakening, as represented by our model, is a truly brittle-ductile transition process in that it occurs under conditions where pressure solution is too slow to accommodate shear without dilatation (velocity-strengthening ductile deformation) yet fast enough to influence frictional behavior by moderating the amount of dilatation that can occur at a given slip velocity to values below ϕ_c .

9. Conclusions

In this study, we aimed to derive a microphysical model to explain the steady state frictional behavior of illite-quartz gouges, based on the approach of *Bos and Spiers* [2001] and *Niemeijer and Spiers* [2005, 2007], but accounting for the matrix supported nature of phyllosilicate-rich gouge. We assumed deformation to be controlled by frictional slip on the phyllosilicates and thermally activated deformation of the quartz clasts. Although the microstructures of deformed illite-quartz mixtures suggested an important role of SCC, reducing the grain size at least initially, the poorly constrained nature of SCC led us to implement pressure solution as the thermally activated deformation mechanism in our model—with the objective of evaluating the potential role of this process in controlling gouge friction. We conclude the following:

1. Incorporating pressure solution plus rate-independent phyllosilicate friction in the model explains many of the main experimental observations on illite-quartz gouges.
2. The trends in steady state friction coefficient μ predicted by the microphysical model, like the experiments reported by *Den Hartog and Spiers* [2013], show an increase in μ with increasing T and little effect of σ_n^{eff} , across much of the range of conditions explored experimentally.
3. A predicted decrease in μ with increasing T at high temperatures and low slip rates, and a predicted decrease in μ with increasing σ_n^{eff} at high temperature, do not agree with the experimental results, but are likely related to the discontinuous transition from non-dilatant to dilatant behavior, which characterizes the model but in reality is expected to be smoothed out.
4. The model predicts zero effect of quartz fraction (f_{qtz}) on μ at low temperatures, whereas experiments on phyllosilicate-quartz gouges show a clear increase in μ with f_{qtz} . This points to a critical state in real gouges

whereby a rate-independent balance between pore opening and closure is achieved at lower porosities than the zero overlap clast configuration characterizing the model.

5. Predicted ($a-b$) versus T profiles shows three regimes of ($a-b$), with velocity weakening at intermediate temperatures, consistent with the experimental observations on illite-quartz mixtures. These regimes are predicted to shift toward higher temperatures with increasing slip rate and decreasing effective normal stress, again in agreement with experiments. The velocity-weakening regime in the predicted ($a-b$) versus T profile shows similar ($a-b$) magnitudes but is displaced toward higher temperatures compared with the experimental results. This may imply localized shear deformation of the aggregate or a velocity-dependence of $\bar{\mu}$ for pure illite, as seen in pure muscovite.
6. The model predicts velocity-neutral behavior at low temperatures, whereas the experiments showed velocity-strengthening behavior. This difference can be explained by the fact that pure phyllosilicates show velocity-strengthening behavior, as opposed to rate-independent phyllosilicate friction behavior assumed in the model.
7. Velocity strengthening predicted at high temperature is in agreement with experimental observations on illite-quartz gouge. The predicted magnitude of ($a-b$), however, is too high, possibly related to the discontinuous transition from non-dilatant to dilatant behavior in the model, which in reality is expected to be smoothed out.
8. Application of the model to slip rates believed to be relevant for earthquake nucleation (10^{-9} – 10^{-7} m/s), but using otherwise equivalent conditions to the experiments on illite-quartz gouge, predicts velocity weakening starting at 140°C, which is in good agreement with the temperature associated with the updip seismogenic limit on subduction megathrusts. Prediction of the temperature at which a switch occurs back to velocity strengthening at higher temperature, i.e., the equivalent of the down-dip seismogenic limit, requires formulation of a microphysical model for muscovite-quartz friction.
9. The similarity between the three regimes of ($a-b$) behavior versus temperature predicted by the model, seen in experiments, and implied by the sequence of aseismic-seismic-aseismic behavior on natural metapelite-dominated megathrusts, suggests that seismogenesis is related to velocity weakening caused by the competition between gouge compaction involving pressure solution of quartz clasts and dilatant frictional slip on the intervening phyllosilicates. Thermally activated pressure solution produces velocity weakening and seismogenesis in the brittle-ductile transition, i.e., under conditions where the creep process is fast enough to moderate “brittle” dilatation to remain at subcritical porosity values but too slow to allow ductile shear of the clast phase.

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