RESEARCH ARTICLE

10.1029/2018JB016053

Localized Slip and Associated Fluidized Structures Record Seismic Slip in Clay-Rich Fault Gouge

Key Points:

- Paired localized slip zones and random fabric gouge record seismic slip at low effective normal stress in clay-rich fault gouge
- Structures and dissipated energy are consistent with thermal pressurization in wet gouge
- Structures and dissipated energy are consistent with thermal decomposition and cataclasis in dry gouge

Correspondence to:

M. E. French,
mefrench@rice.edu

Citation:

French, M. E., & Chester, J. S. (2018). Localized slip and associated fluidized structures record seismic slip in clay-rich fault gouge. *Journal of Geophysical Research: Solid Earth*, 123. <https://doi.org/10.1029/2018JB016053>

Received 2 MAY 2018

Accepted 24 SEP 2018

Accepted article online 27 SEP 2018

Melodie E. French¹ and **Judith S. Chester²**

¹Department of Earth, Environmental, and Planetary Sciences, William Marsh Rice University, Houston, TX, USA, ²Department of Geology and Geophysics, Center for Tectonophysics, Texas A&M University, College Station, TX, USA

Abstract Fault rocks can weaken dramatically with increasing slip rate, which results in localization of slip and earthquakes. Exhumed fault zones and fault rocks deformed at seismic rates in the laboratory both show that deformation can become extremely localized to zones less than or equal to millimeters thick. However, localization can occur during aseismic slip, so evidence of localization cannot necessarily be interpreted as having occurred coseismically. Dynamic weakening that occurs during earthquakes is the result of processes that are unique to seismic slip rates, and previous results from carbonates show that these processes produce unique microstructures. We evaluate whether coseismic deformation at low normal stress produces unique structures within the localized slip zones and adjacent gouge that develop in clay-rich gouge from the Central Deforming Zone of the San Andreas fault. We measured the thickness and orientations of localized slip zones and their internal lamina, particle orientations, and particle size distributions of gouge sheared from 0.35 to 1.3 m/s velocity, up to 25-m displacement, and 1-MPa normal stress, under water-wet and room-dry conditions. We find that the thicknesses of localized slip zones and their internal laminae are consistent with numerical formulations for thermal pressurization in wet gouge and both thermal decomposition and cataclastic deformation in dry gouge. Localized zones form coincident with a zone of fluidized gouge that accommodates at most 10% of the shear strain. We conclude that the combined occurrence of foliated localized shear zones with a zone of fluidized gouge may provide a record of seismicity in clay-rich gouges.

1. Introduction

Many exhumed fault zones have a very thin (less than or equal to centimeter-scale) ultracataclasite or gouge layers at their cores, which is taken as evidence that fault displacement was accommodated within a very narrow zone of localized slip (e.g., Chester & Chester, 1998; Sibson, 2003; Smith et al., 2011; Wibberley & Shimamoto, 2005). The localization of deformation is identified on the basis of intense grain size reduction and offset markers and by the lack of evidence for large shear strains in the surrounding damage zone rock (Chester & Logan, 1986; Chester et al., 2005; Di Toro & Pennacchioni, 2005). In contrast, in exhumed fault zones that are rich in phyllosilicates, the highest strains can be distributed over much wider zones of deformation (Faulkner et al., 2003; Moore & Byrne, 1987; Rowe et al., 2013). For fault deformation to localize to a narrow zone, the strength of the rock must decrease with increasing slip, slip rate, or both (Beeler et al., 1996; Ben-Zion & Sammis, 2003; Scruggs & Tullis, 1998). Because the weakening of fault rock with slip or slip rate is also a prerequisite for earthquakes, localization is expected to occur during coseismic fault slip (Dieterich, 1978). As a result, localized slip zones potentially provide an equivocal record of seismicity (Cowan, 1999). Furthermore, whereas faults that accommodate all offset during earthquakes are expected to develop at least one narrow high-strain zone, faults that exhibit periods of both creep and seismogenesis likely record a more complex history of strain partitioning that includes periods of distributed and localized deformation (e.g., Rowe et al., 2011; Ujiie et al., 2007). As a result, it can be difficult to recognize evidence for both localization and seismic slip in clay-rich faults.

Detailed structural analyses of fault gouges sheared at seismic rates in the laboratory reveal structures that potentially develop only at seismic slip rates, as they have not been reported in experiments conducted at subseismic rates (Kitajima et al., 2010; Mizoguchi et al., 2009; Rempe et al., 2014; Smith et al., 2017). Slip consistently localizes to zones 1- to 100- μ m thick concurrent with both apparent slip and rate weakening, but the character of these slip zones varies with lithology, pore fluid content, and potentially normal stress

(Aretusini et al., 2017; Brantut et al., 2008; Bullock et al., 2015; Mizoguchi et al., 2009; Smith et al., 2011). The mechanisms that result in localization at seismic rates are thought to be controlled by shear heating and associated thermal weakening processes (Rice et al., 2014; Platt et al., 2014, 2015), and these processes can produce unique microstructures (Brantut & Sulem, 2012). Recent work provides quantitative measurements of foliation orientations and grain size distributions that develop during seismic slip in carbonates (e.g., Rempe et al., 2017; Smith et al., 2017) and a wealth of qualitative observations on grain size and foliation orientations for siliciclastic and clay-rich lithologies (Aretusini et al., 2017; Kitajima et al., 2010; Mizoguchi et al., 2009; Ujiie & Tsutsumi, 2010). Our goals are (1) to provide quantitative measures of the microstructures that develop during seismic slip of clay-rich gouge at low normal stress, (2) to determine the processes that lead to the development of these microstructures, and (3) to evaluate if structures in exhumed localized clay-rich slip zones can be interpreted as having developed at seismic rates.

We previously documented the development of localized slip zones and an adjacent region of random fabric in experiments conducted at coseismic rates on smectite-rich gouge from the Central Deforming Zone (CDZ) of the San Andreas fault (SAF; French et al., 2014). Here we quantify the thickness of these localized zones and compare these data to theoretical models for slip zone thickness during thermal pore fluid pressurization and thermal decomposition weakening (Platt et al., 2014, 2015). We present the particle sizes, size distributions, and preferred orientations within the localized shear domain and the adjacent random fabric domain to constrain deformation processes and strain partitioning within seismically shearing clay gouge. Finally, we interpret these structures in the context of our previously reported measurements of mechanical energy dissipation and thermal heating (French et al., 2014) to constrain the coseismic deformation processes that lead to localization and discuss how these processes may be interpreted from the rock record.

The CDZ gouge recovered during the San Andreas Fault Observatory at Depth program (Hole G, Run 4, section 4, 9 cm) is composed of a phyllosilicate matrix that is dominantly saponite with additional chlorite, illite, and chrysotile and clasts derived from serpentinite and sedimentary protoliths (Ikari et al., 2016; Moore, 2014; Schleicher et al., 2009, 2012). Using optical microscopy, we determined our CDZ sample to be 90–95 vol % phyllosilicate matrix and 5–10 vol % clasts that are primarily serpentinite, quartz, calcite, and oxides (French et al., 2014), consistent with other samples of CDZ gouge whose composition is reported (Bradbury et al., 2011; Ikari et al., 2016; Moore, 2014). Our CDZ gouge was taken from the same block studied by Coble et al. (2014) and French et al. (2015) and has an X-ray diffraction profile identical to the pattern reported by Lockner et al. (2011). The intact CDZ gouge has a scaly fabric with shiny striated shear surfaces and is semicohesive (Schleicher et al., 2010); it holds its shape but can be broken apart by hand.

2. Strength and Thermal Evolution of CDZ Gouge

Previously, we measured the constitutive behavior of fault gouge from the CDZ at seismic slip rates (French et al., 2014) and at extremely low slip rates (French et al., 2015) to constrain the potential for seismic slip along the creeping segment of the SAF. In the 2014 study, we related the strength of the fault rock as a function of slip and slip rate to the thermal evolution of the gouge during shear and qualitatively described the concurrent evolution of microstructures. In this paper we quantify key microstructural characteristics that reflect coseismic slip of the gouge in these experiments.

We sheared ~1.2-mm-thick layers of CDZ gouge between dolerite host blocks 25 mm in diameter in a rotary shear configuration (French et al., 2014; Figure 1a). The gouge was prepared by gently flaking it with a razor to preserve the natural microfabric and then sieving it and retaining fragments <634 μm . The gouge was sheared either at room humidity in a *room-dry* condition or in a *water-wet* condition by adding 0.3 g of deionized water to the gouge to form a paste prior to sample assembly. During rotary shear, one host block is rotated at a constant angular velocity, while the other host block does not rotate but moves parallel to its axis to maintain a constant normal force across the gouge layer. In this geometry, slip velocity and total displacement vary from 0 at the sample center to a maximum at the sample end (e.g., Kitajima et al., 2010). Following convention and for simplicity, we refer to the *representative* deformation conditions at two thirds the sample radius (8.3 mm) unless otherwise specified (Shimamoto & Tsutsumi, 1994). The strength of the CDZ gouge was measured in 30 experiments conducted at 1-MPa total normal stress, representative velocities of $v_r = 0.1, 0.35, 0.7,$ and 1.3 m/s, and representative displacements d_r from 1 to 25 m. Here we analyze eight of these samples in detail (Table 1).

During shear, the CDZ gouge initially strengthens over 1.5-m representative displacement concomitant with

Figure 1. (a) Sample configuration showing the gouge layer (brown) sandwiched between dolerite host blocks. The rotating host block has fixed axial position, and the stationary host block cannot rotate but can move axially. The gouge is contained with a Teflon ring. (b) Radial and tangential thin sections were prepared from gouge layers. Radial sections and tangential sections show structures that develop normal and parallel to the slip vector, respectively. Modified Figure 2 from French et al. (2014).

the formation of a distinct foliation throughout the gouge. This foliated zone, referred to as Unit 2, develops from the precompacted and slightly sheared gouge defined as Unit 1 (Figures 2, 3a, and 3b). Unit 2 develops across the entire gouge layer and has a composite planar foliation with shear surfaces in P and Y orientations, following the classification of Logan et al. (1979) (Figure 3b). At representative slip rates of 0.35 m/s and higher, the gouge layer weakens to an apparent steady state strength at ~ 10 -m representative displacement. During weakening, the foliated unit (Unit 2) is progressively destroyed and overprinted by a localized zone of microfoliated gouge, called Unit 4, and an adjacent zone of lighter colored and unfoliated gouge, called Unit 3 (Figures 2 and 3c–3e). During the development and shearing of Units 3 and 4, the gouge layer dilates and the magnitude of dilation correlates with slip velocity and temperature (Figures 2b and 2b). This general correlation between slip weakening, dilatancy, and fabric development occurs under water-wet and room-dry conditions (Figure 2). The steady state strength decreases with increasing slip rate, and the gouge is weaker when water-wet relative to the dry gouge under all deformation conditions.

While the character of the microstructurally defined Units 1, 2, and 3 are similar under wet and dry conditions under the petrographic microscope, the character of Unit 4 differs significantly (Figures 3c–3e). We refer to Unit 4 in water-wet and room-dry gouges as 4-W and 4-D, respectively. Unit 4-W is dark brown under plane- and cross-polarized light, has a foliation that is similar to Unit 2, and a thickness of approximately 100 μm (Figure 3c and

Table 1). Unit 4-D is lighter in color, has uniform birefringence under cross-polarized light, and thickness that decreases 2 orders of magnitude from 500 to 5 μm as the representative slip rate increases from 0.35 to 1.3 m/s (Figures 3d and 3e and Table 1).

On the basis of gouge strength, thermal evolution, and microstructural development, we previously inferred that displacement and rate weakening of the wet gouge result from slip along the clay foliation of Unit 4-W assisted by shear-heating pressurization of pore fluid (French et al., 2014; Figures 2 and 3). We hypothesized that Unit 4-D deforms by more extreme grain size reduction and possible clay dehydration. On the basis of fabric development, injection structures, dilatant behavior, and thermal evolution, we interpreted Unit 3 to be fluidized gouge that developed during the thermal expansion of pore fluids and adsorbed water and dehydration of interlayer water (French et al., 2014). Fluidization refers to a state in which the mean free path between particles is larger than their diameter, resulting in a fluid-like behavior (e.g., Monzawa & Otsuki, 2003). These inferences are supported by observations that the evolution of strength and fabric coincide with a transition

Table 1
List of Experiments and Conditions

Experiment #		v_r^a (m/s)	Units present	w_{U4}^b (μm)	τ_{ss}^c (MPa)	E_l^d (J)	δw^e (mm)	$w_{\text{calc}} \text{ TP}^f$ (μm)	$w_{\text{calc}} \text{ TD}^g$ (μm)
2403	Dry	1.30	2, 3, 4	5	0.25	970	0.063		4.2
2421	Dry	0.70	2, 3, 4	15	0.23	861	0.05		6.2
2420	Dry	0.35	3, 4	500	0.35	1,116	0.045		12.5
2419	Dry	0.10	n/a	n/a	0.51	531	−0.1		50
2404	Wet	1.30	2, 3, 4	85	0.15	390	0.03	45	5.7
2410	Wet	0.70	1, 2, 3, 4	90	0.20	440	0.063	66	8.5
2409	Wet	0.35	1, 2, 3, 4	111	0.30	729	0.02	144	17
2408	Wet	0.10	1, 2	n/a	0.37	875	−0.03	2,050	68

^a v_r is representative velocity. ^b w_{U4} is the maximum thickness of Unit 4 measured from each sample. ^c w_{U4} is the maximum thickness of Unit 4 measured from each sample. ^d E_l is the latent heat dissipated during the experiment determined in French et al. (2014). ^e δw is axial shortening at the end of the experiment. ^f $w_{\text{calc}} \text{ TP}$ is the calculated localized shear zone thickness during thermal pressurization (wet gouge only) based on Platt et al. (2014). ^g $w_{\text{calc}} \text{ TD}$ is the calculated thickness during thermal decomposition based on Platt et al. (2015).

Figure 2. (a) Shear strength evolution, and (b) axial shortening (solid lines) and calculated temperature (dashed lines) of water-wet gouge with representative displacement. (c) Shear strength evolution, and (d) axial shortening (solid lines) and calculated temperature (dashed lines) of room-dry gouge with representative displacement. The temperature shown is the maximum temperature in the gouge layer as calculated by French et al. (2014). Positive and negative axial shortening indicate gouge compaction and dilation, respectively. Colors indicate the approximate range of displacements over which each unit or set of units is identified as accommodating deformation in French et al. (2014).

from gouge compaction during slip-hardening to dilation and a gradual temperature increase during slip weakening (Figures 2b and 2d).

3. Microscopy

To quantify the microscale fabric of the different units, we measured the thickness and orientation of different sets of lamina, particle orientations, and particle size distributions in Units 1, 3, 4-W, and 4-D. We use the term *particle* to refer to clasts, clay grains, and aggregates of clay collectively. The term *clasts* describes both single grains of nonphyllosilicate minerals and composites of multiple grains, including serpentinite fragments. We use the term *matrix particles* to refer to particles too small to resolve using scanning electron microscope (SEM) imaging; these consist of clay grains and aggregates of clay grains.

Unit 1 was investigated from a gouge layer that was saturated and precompacted at 1-MPa normal stress but not sheared (Figure 3a). Fabric elements of Unit 3 were measured from sample #2420 over 3 orders of magnitude in spatial scale. We checked that these results are representative of Unit 3 by selectively measuring the same fabric elements from a sample deformed under wet conditions and a sample deformed dry at a different slip rate (Table 2). The thicknesses of Units 4-W and 4-D were measured from every sample. We measured internal fabric elements of Unit 4-D from sample #2420, which is the only sample in which this unit is abundant enough to measure a statistically significant number of particles and laminae.

Polished 30- μm -thick sections were studied using optical microscopes and SEMs at magnifications from 10 $\times$ to 10,000 $\times$. The SEM was used to collect back-scattered electron (BSE) images at an excitation voltage of 20 keV and working distance of 10.0 mm and at magnifications from 150 $\times$ to 10,000 $\times$. Bright field transmission electron microscope images were collected from sections mounted to Cu grids and Ar⁺ ion milled. Transmission

Figure 3. The deformation units defined by French et al. (2014) shown in tangential sections. (a–c) Plane-polarized light images and (d, e) cross-polarized light images. The units and their significance are described in the text.

Table 2*Image Details and Microstructural Parameters for Samples Analyzed for Clast and Clay Size Distributions and Orientations*

Experiment #	Image type	Scale	Image width (μm)	Image height (μm)	Number of grains	a^a	$\hat{V}^a$	$\bar{a}^b$
Unit 3, tangential sections								
2420	PM	10×	4,992	1,766	105		19	0.11
2420	BSE	150×	826	598	1,336		−15	0.16
2420	BSE	1,000×	311	189	1,959		−16	0.08
				#2420 Average	3,400	2.2	165	0.11
2409	PM	10×	1,967	122	109	2.2	−2.9	0.22
2403	BSE	150×	2,994	1,182	422	2.1	−3.8	0.16
2403	BSE	1,500×	280	177	343	2.2	0	0.11
Unit 3, radial sections								
2420	BSE	100×	1,456	1,340	810		−4	0.17
2420	BSE	1,000×	153	131	1,500		5	0.18
2420	BSE	2,500×	54	47	323		−19	0.28
				#2420 Average	2,633	2.4	178	0.18
2420	TEM	5,000×			1,654	2.2	12	0.07
2409	BSE	150×	1,036	874	488	2.0	−4.6	0.27
2403	BSE	150×	2,772	1,184	660	2.0	−2.9	0.20
Unit 4, tangential sections								
2420	BSE	100×	1,515	254	51		−23	0.45
2420	BSE	300×	379	207	873		13	0.63
				#2420 > 3 μm Average	1,447	4.0	13	0.58
2420	BSE	2,500×	123	78	523	0.9	17	0.54
Unit 4, radial sections								
2420	BSE	200×	552	108	362		−1	0.69
2420	BSE	2,500×	177	82	559		1.7	0.64
				#2420 > 3 μm Average	2,264	6.5	1.7	0.64
2420	BSE	10,000×	38	24	127	0.97	4.0	0.6
2420	TEM	5,000×			1,216	2.6	−14	0.39
Unsheared/Unit 1								
2441	PM	10×			963	1.8	0.3	0.28

Note. PM = Petrographic microscope; BSE = back-scattered electron; TEM = transmission electron microscope.

^a a is the PSD exponent. ^b $\hat{V}$ is the mean orientation of the long axes of clasts and clay grains. ^c $\bar{a}$ is the strength of $\hat{V}$.

electron microscope (TEM) operating conditions were 200 keV and 110 μA , and images were collected at a magnification of 5,000×

Fabric elements were quantified from both radial and tangential sections (Figure 1b). Radial sections show structures perpendicular to the slip vector, and tangential sections show structures that are parallel to the slip vector at the thin section center and become increasingly oblique to the slip vector with increasing distance from the center (Kitajima et al., 2010). Structures in tangential sections were analyzed within ± 2.5 mm of the thin section center where the slip vector is subparallel to the image plane.

Particle maps were made by segmenting grayscale images or by digitally tracing them (Heilbronner & Keulen, 2006). Laminae were digitally traced from optical microscope and BSE images. Particle sizes and orientations, and the orientation of lamina boundaries were measured from the maps using the ImageJ software. Particle orientation is reported as the angle of the long axis of a best fit ellipse with respect to the shear plane on a scale from -90° to $+90^\circ$. The size of particles is reported as the equivalent diameter of a circle having a cross-sectional area equivalent to that of the particle (e.g., Sammis et al., 1987).

To quantify the orientations of particles, we report a mean vector orientation and strength for each image. The mean vector orientation for N particles is determined by

$$\hat{V} = \frac{1}{2} \tan^{-1} \left(\frac{\sum_{i=1}^N \sin 2\theta_i}{\sum_{i=1}^N \cos 2\theta_i} \right) \quad (1)$$

where θ is the orientation of the long axis of particle i (Agterberg, 1974). The vector strength is given by

$$\bar{a} = \frac{1}{N} \left[\left(\sum_{i=1}^N \sin 2\theta_i \right)^2 + \left(\sum_{i=1}^N \cos 2\theta_i \right)^2 \right]^{1/2} \quad (2)$$

where $\bar{a} = 1$ when all long axes are parallel and $\bar{a} = 0$ when long axes are uniformly distributed (Agterberg, 1974). Cladouhos (1999) found that a random distribution of particles has > 95% chance of $\bar{a} < 0.17$. Thus, we consider distributions with $\bar{a} > 0.17$ to be a statistically significant preferred orientation.

Two-dimensional particle size distributions are reported as the cumulative frequency of particles larger than a given equivalent particle size. At each image magnification, particles are binned by size with 20 bins per unit of magnitude particle size. Particle frequency, $N(D)$, is measured by summing the number of particles in bins greater than equivalent diameter, D , divided by the total image area at that magnification. Consistent with most analyses of particle size distribution (PSDs) (e.g., Keulen et al., 2007; Sammis et al., 1987), we test for the power law relation $N(D) = cD^{-a}$ in 2-D by presenting the data in log-log space (Turcotte, 1997). Larger values of a indicate more uniform particle sizes.

4. Results

4.1. Compacted Gouge: Unit 1

Unit 1, the compacted but unsheared CDZ gouge, consists of clasts within a phyllosilicate matrix (Figure 3a). Polymineralic clasts are dominantly serpentinite fragments, and monomineralic clasts are dominantly quartz and oxides. The long axes of clasts in Unit 1 have a weak preferred orientation subparallel to the shear plane ($\bar{a} = 0.39$; Figure 4a). The clay matrix does not exhibit a foliation or other evidence of localized shear surfaces, which are characteristic of Unit 2 and develop during low-displacement strain hardening. The PSD power law exponent for clasts in the compacted gouge (Unit 1) is $a = 1.8$ (Figures 5a–5d and Table 2).

4.2. Localized Zone of Microfoliated Gouge: Units 4-W and 4-D

The maximum thickness of Unit 4-W in the water-wet gouge increases slightly with decreasing representative velocity from 85 μm at 1.3 m/s to 111 μm at 0.35 m/s. The foliation of the localized zone of shear of Unit 4-W is defined by layers subparallel to the macroscopic shear plane that are highlighted by epoxy-filled partings between the extremely fine-grained laminae (Figure 6a). We infer that the partings represent unloading cracks and that the unit is friable, as suggested by the presence of epoxy-filled holes where clasts were plucked during sample preparation. The loss of clasts prohibited quantitative analyses of clast characteristics in this unit.

The maximum thickness of Unit 4-D increases slightly from 5 to 15 μm between 1.3 and 0.7 m/s and then to 500 μm at 0.35 m/s. Unit 4-D contains distinct laminae of fine-grained gouge, separated by sharp boundaries and sometimes unloading cracks (Figures 6b and 7). The gouge within a lamina is often darkest toward the rotating block and becomes brightest toward stationary block in BSE images, indicating a decrease in porosity potentially enhanced by reduced grain size (Figure 7). The matrix of Unit 4-D consists of lens-shaped particles that are composed of one or more clay grains and distorted circular aggregates of clay, similar to aggregates reported in similarly deformed talc gouges (Figure 8a; Boutareaud et al., 2012). The matrix also contains a very fine-grained material, presumably composed of clay, with particle sizes and shapes that cannot be resolved at the magnification of these images (Figure 8a). Similar nanoparticles of smectite were produced during other high-speed friction tests (Aretusini et al., 2017). Clasts of serpentinite, quartz, and oxides are distributed throughout Unit 4-D and are higher density than the matrix (Figure 7). There is no evidence that clasts were plucked during thin section preparation, presumably because Unit 4-D is much less friable than Unit 4-W.

Lamina boundaries occur in two orientations that define an asymmetry in tangential sections (Figures 7a and 9a). One set of boundaries is oriented approximately -10° to $+5^\circ$ to the shear plane; these boundaries are the most abundant and the longest (Figure 9a). A second set of lamina boundaries consists of shorter segments

Figure 4. Histograms showing the orientations of clasts measured from petrographic and back-scattered electron images and matrix particles measured from transmission electron microscope images. Measurements at different magnifications are distinguished by color, and histograms with measurements at multiple magnifications show cumulative data. The n is the number of measurements at a given magnification. (a) Clasts in a sample that was precompactified but not sheared (#2441). Measurements in (b)–(g) were made on sample #2420, and measurements from other samples are similar and are provided in Table 2. (b) Clasts in a tangential section of Unit 4-D. (c) Clasts in a radial section of Unit 4-D. (d) Matrix particles from a radial section of Unit 4-D. (e) Clasts in a tangential section of Unit 3. (f) Clasts in a radial section of Unit 3. (g) Matrix particles in a radial section of Unit 3.

Figure 5. Particle size distributions of (a–d) clasts measured from polarized light images and back-scattered electron images, and (e, f) matrix particles measured from transmission electron microscope images of sample #2420. Comparison with additional samples is provided in Table 2. The particle size distribution of clasts in a sample that was compacted but not sheared (Unit 1, sample #2441) is shown in light gray in (a)–(d). Data become nonlinear at the high and low cumulative frequency limits due to sampling bias and finite size of images. Power law exponent fits were determined only for cases where the log-log relation is linear over at least half an order of magnitude.

oriented 10° to 20° to the shear plane. This latter set connects but does not crosscut the primary set. In radial sections, lamina boundaries are symmetric and subparallel (-20° to $+20^\circ$) to the shear plane (Figures 7b, 9c, and 9d). The two sets of lamina boundaries are oriented similar to *R* and *P* type shear surfaces (Logan et al., 1979). Measured normal to the shear plane, the average spacing between lamina boundaries is 7.7 and $5.1 \mu\text{m}$ in radial and tangential sections, respectively.

Both clasts and matrix particles have preferred orientations in Unit 4-D (Figures 4b–4d). In tangential sections, the long axes of clasts are oriented between subparallel and 25° to the shear boundary in a *P* orientation

Figure 6. Backscattered electron images of radial sections showing the slip-normal structure of Units 3 and 4 in gouge sheared (a) wet ($v_r = 0.35 \text{ m/s}$, #2409) and (b) dry ($v_r = 0.35 \text{ m/s}$, #2420). The black dashed box in (b) is the field of view in Figure 7b. The rotating host blocks are toward the bottom of the images. Green lines show boundaries between Units 3 and 4. Red bracket in (b) shows region of large and variable clast size.

Figure 7. Backscattered electron images of Unit 4-D ($v_r = 0.35$ m/s, #2420) in (a) tangential and (b) radial sections. The rotating host blocks are toward the bottom of the images. Red lines outline one lamina, which shows an increase in brightness upward away from the rotating block. In (a), lamina from the dominant set are indicated with solid red lines and those from the secondary set are indicated with dashed red lines. Black arrows point to clasts, and white arrows point to unloading cracks.

($\hat{V}_{\text{avg}} = 13^\circ$ and $\bar{a}_{\text{avg}} = 0.58$; Figure 4b and Table 2). In radial sections, the long axes of clasts have preferred orientations subparallel to and symmetric about the shear plane ($\hat{V}_{\text{avg}} = 1.7^\circ$ and $\bar{a}_{\text{avg}} = 0.64$). Matrix particles in radial sections have a weaker, but statistically significant, preferred orientation similar to R surfaces ($\hat{V} = -14^\circ$ and $\bar{a}_{\text{avg}} = 0.39$; Figures 4c and 4d and Table 2).

Clasts in Unit 4-D appear relatively uniform in size, except near the transition between Units 4-D and 3 (Figure 6b). There is a change in PSD of the clasts at a size of $3 \mu\text{m}$, and it occurs in both radial and tangential sections (Figures 5a and 5b). The distribution of clasts larger than $3 \mu\text{m}$ can be fit with power law exponents of 4.0 and 6.5 in tangential and radial sections, respectively. This difference in PSD is consistent with clasts having long axes that are preferentially oriented in the plane containing the slip vector in three dimensions. The PSD of clasts smaller than $\sim 3 \mu\text{m}$ are best fit with a power law exponent of 0.9 and 1.0 in tangential and radial sections, respectively, indicating clast sizes are much less uniform at small sizes. In contrast, matrix particles, which were measured over the size range 0.02 to $2 \mu\text{m}$, have a PSD of 2.7 in a radial section, which is larger than the PSD of similarly sized clasts and smaller than that of larger clasts (Figure 5e).

4.3. Fabric of Fluidized Gouge: Unit 3

In contrast to Unit 4, Unit 3 does not display a foliation at any scale of observation (Figures 3c–3e, 6, and 8b). Like Unit 4-D, the matrix contains lens-shaped clay grains and grain aggregates (Figure 8b). However, the very fine-grained material that comprises a significant proportion of the matrix in Unit 4-D is much less abundant in Unit 3. Unit 3 also appears to be more porous than Unit 4-D on the basis of greater light

Figure 8. Transmission electron microscope brightfield images of (a) Unit 4-D and (b) Unit 3 in slip-normal sections ($v_r = 0.35$ m/s, #2420, Table 2) illustrating matrix particles described in the text. The rotating host block is toward the bottom of the images. Red arrows indicate lens-shaped clay particles, black arrows indicate clay aggregates, and the blue arrow indicates very fine grained material.

Figure 9. The frequency (a and c) and length (b and d) of surfaces separating lamina in Unit 4-D (Sample #2420, Table 2). (a) Histogram of the frequency of lamina boundaries with orientation in a tangential section showing two populations at approximately $+10^\circ$ and -5° to the shear plane and (b) scatterplots of the trace lengths for the same boundaries in (a). (c) Histogram of the frequency of lamina boundaries with orientation in a radial section showing a single population symmetric about the shear plane (b) scatterplots of the trace lengths of the same boundaries as in (d). The n is the number of boundaries measured.

transmission in plane-polarized optical images (Figure 3c) and the structure of grains in TEM images (Figure 8). At the scale of our TEM measurements ($5,000\times$), the PSD of matrix particles in Unit 3 and Unit 4-D is very similar (Figure 5f). The matrix particles in the radial TEM section of Unit 3 analyzed does not have a statistically significant long-axis preferred orientation ($\bar{V}_{\text{avg}} = 12^\circ$ and $\bar{a}_{\text{avg}} = 0.06$; Figure 4g).

Clasts in Unit 3 either do not have a statistically significant preferred orientation or is weak, in both tangential and radial sections (Figures 4e and 4f). In cases where there is a weak preferred orientation, it is subparallel to the shear plane (Figure 4f and Table 2). In general, clasts in Unit 3 are smaller than those in Unit 1 and larger than those in Unit 4 (Figure 6). As in Unit 4-D, the PSD of Unit 3 clasts measured from the same sample indicates that the sizes are slightly more uniform in slip-normal (e.g., $a = 2.4$ for sample #2420) planes than in slip-parallel (e.g., $a = 2.2$ for sample #2420) planes (Figures 5c and 5d). Clast sizes in Unit 3 are more uniform than those in Unit 1 ($a = 1.8$) and much less uniform than those in Unit 4-D (Figures 5c and 5d). The PSD of matrix particles ($a = 2.2$) is similar to that of the clasts.

5. Discussion

The development of a localized zone of foliated gouge (Unit 4) with an adjacent fluidized zone of gouge (Unit 3) during weakening at seismic slip rates is described for a number of lithologies (e.g., Brantut et al., 2008; Kitajima et al., 2010; Mizoguchi et al., 2009; Rempe et al., 2014; Yao et al., 2013), and we argue that the co-occurrence of these units records the processes that occur during seismic slip under low effective normal stresses. We use these new structural measurements in Units 3 and 4, combined with our previously reported mechanical data and thermal evolution, to (1) evaluate strain partitioning between these units at

seismic slip rates and (2) test existing models for coseismic weakening in fault gouge. We independently investigate the processes that may lead to the fabrics developed in Units 3 and 4 and their hypothesized shear strengths. These analyses are used to constrain the processes that accommodate displacement and lead to dynamic weakening in clay-rich fault gouge at low normal stresses and to provide criteria for identifying these processes in the geologic record.

The kinematics of strain in a deforming gouge unit can be characterized by the parameter $s = \frac{\dot{\epsilon}}{\dot{\gamma}}$, which represents the relative rates of pure shear strain, $\dot{\epsilon}$, and simple shear strain, $\dot{\gamma}$ (Cladouhos, 1999; Ghosh & Ramberg, 1976). Under our experimental boundary conditions, any pure shear strain component occurs either by dilation or compaction of the gouge unit or by migration of the boundary of that unit normal to the shear boundary, assuming that radial strain is negligible. Given that the magnitudes of gouge dilation and compaction are on the order of ~ 0.1 mm and the maximum unit boundary migration is the layer thickness and ~ 0.1 to 0.9 mm, there is at most ~ 1 mm of pure shear component per ~ 10 m of slip in any unit corresponding to $s = 1 \times 10^{-4}$ (Figures 2b and 2d). It follows that the entire gouge layer is dominated by simple shear kinematics. We conclude below that Unit 4 accommodates most of the shear strain, so it too is dominated by simple shear kinematics.

5.1. Comminution

The PSD of clasts in the compacted gouge is expected to be representative of the in situ CDZ gouge over the range of clast sizes in our sample, because we disaggregated it without grinding during sample preparation. The starting material PSD of $a = 1.8$ is close to the predicted value of 1.6 for constrained comminution (Abe & Mair, 2005; Sammis et al., 1987) and similar to PSDs reported for fault rock deformed in situ and experimentally at subseismic strain rates and low strain (An & Sammis, 1994; Keulen et al., 2007). The smaller clast size and higher PSD of clasts in Unit 3 ($a = 2.0$ and 2.4 in tangential and radial sections of #2420, respectively) relative to the starting material indicates preferential fracture of large clasts throughout development of the distributed foliation (Unit 2) and then subsequent overprinting by Unit 3. Similar PSDs have been reported for both seismic strain rates and high strains in fault rocks, but these measurements are often made outside the localized slip zones in these rocks where total strain and strain rate may have been low, as we infer to be the case for Unit 3 (Hadizadeh & Johnson, 2003; Monzawa & Otsuki, 2003; Stunitz et al., 2010).

The smaller clast and matrix particle sizes of Unit 4-D relative to Unit 3 indicates that Unit 4-D accommodates higher shear strains than Unit 3 (Figure 4). The larger PSD exponent of clasts in Unit 4-D ($a = 4.0$ and 6.5), relative to Unit 3, is also consistent with this interpretation, as the PSD exponent has been shown to become larger with increasing strain and decreasing mean particle size (An & Sammis, 1994; Marone & Scholz, 1989; Storti et al., 2003). In contrast to the clast data, the PSDs of matrix particles are very close in Units 3 and 4-D despite evidence of size reduction (Figure 4e), implying that different processes control the PSD exponents of clasts and matrix particles.

The PSD exponents of clasts in Unit 4-D are greater than most previous studies have reported. Wilson et al. (2005) reported PSD exponents up to $a = 6$, but their rocks were pulverized damage zone rocks that did not accommodate measurable shear strain. The change in power law exponent at a clast size of 1 to $3 \mu\text{m}$ in Unit 4-D occurs at the approximate grinding limit for silicate and carbonate minerals (An & Sammis, 1994; Prasher, 1987), and the power law exponent (~ 1) for the small size fraction is consistent with fragments produced by attrition below the grinding limit (Keulen et al., 2007). The consistency of the result $a \approx 1$ at clast sizes below $1 \mu\text{m}$, in this study and others (Hirose et al., 2012; Keulen et al., 2007; Stunitz et al., 2010), over orders of magnitude in strain rate and strain is striking. This result is particularly notable given the power law exponent for clast diameters greater than $1 \mu\text{m}$ varies from 1.5 to 6 in these same studies. We conclude, therefore, that the high PSD of clasts in Unit 4-D developed by the spalling of small fragments off of clasts as they approached their grinding limit. This resulted in a large clast size fraction with relatively uniform size and a small clast size fraction with a size distribution characteristic of being below the grinding limit.

5.2. Mechanisms of Localization

Slip localizes during deformation that is slip weakening, rate weakening, or a combination of the two, and at steady state the thickness of the localized slip zone reflects an intrinsic length scale of the deformation process. This length scale may be controlled by grain size if the deformation process is granular (Morgens-tern & Tchalenko, 1967; Muhlhaus & Vardoulakis, 1987) or by thermal and hydraulic diffusion if the process is pore fluid pressurization or dehydration weakening (Rice et al., 2014; Platt et al., 2014, 2015). We evaluate the consistency of the thickness and internal structure of Units 4-D and 4-W with these controls.

In fault gouge that has rate-strengthening constitutive properties, like the CDZ gouge (Carpenter et al., 2012; Coble et al., 2014; French et al., 2015; Lockner et al., 2011), coseismic shear heating and resultant thermal pressurization can cause the apparent slip weakening and rate weakening observed in the CDZ gouge (French et al., 2014; Rice, 2006). Such thermally driven coseismic processes can also cause the localization of inherently slip- or rate-strengthening fault rocks to a localized slip zone that has finite thickness (Rice, 2006; Rice et al., 2014). In such a system, thermal weakening is greater in magnitude than the inherent slip or rate strengthening of the fault rock. However, the inherent strengthening processes modulates the thickness of the localized slip zone and prevents localization to a plane (Rice et al., 2014; Platt et al., 2014, 2015).

Motivated by our mechanical data and thermal evolution (Figure 2), we parameterize the models of Platt et al. (2014) for thermal pressurization of gouge sheared water-wet and the model of Platt et al. (2015) for pressurization associated with thermal dehydration of gouge sheared water-wet and room-dry. For thermal pressurization, the thickness of the localized slip zone is expected to depend on the bulk thermal and hydraulic properties, as well as on the shear strength and its rate dependence (Platt et al., 2014; Rice et al., 2014). During thermal dehydration weakening of the smectite in the gouge, the thickness depends on the same parameters, as well as on the kinetics of the reaction (Platt et al., 2015). We use frictional and hydraulic parameters from laboratory studies of the CDZ gouge. The CDZ gouge is inherently rate strengthening at all slip rates at which it has been measured, and we accordingly assume that these properties persist to coseismic rates as well (Carpenter et al., 2012; Coble et al., 2014; French et al., 2015; Lockner et al., 2011). Smectite undergoes two dehydration reactions, the first of which occurs within the temperatures produced during these experiments (Appendix A). The dolerite host blocks have higher thermal diffusivity than the CDZ gouge and tends to lower the temperature of the gouge layer faster than might occur in situ (French et al., 2014; Kitajima et al., 2010). Because the formulations of Platt et al., 2014 (2014, 2015) do not account for this effect, the calculated thicknesses are lower bounds and the effect may explain some of the differences between measured and calculated thickness. Details of these calculations are in Appendix A.

The models predict slip zone thicknesses that decrease from 140 to 45 μm as the slip velocity increases from 0.35 to 1.3 m/s (Table 1). These modeled values are 126% and 53% of the measured thicknesses of Unit 4-W, which decrease from 111 to 85 μm over the same range of representative velocities. The trend of decreasing thickness with increasing slip velocity across the layer is consistent between the modeled and measured values. For gouge sheared at a representative velocity of 0.1 m/s, the model predicts a localized thickness of 1.8 mm, which is greater than the thickness of the sample and consistent with our observation in French et al. (2014) that deformation was distributed across the 1.2-mm thick gouge layer at this condition.

For wet gouge undergoing thermal dehydration, the modeled localized slip zone thickness ranges from 6 μm at a slip velocity of 1.3 m/s to 67 μm at 0.1 m/s, which is 1 to 2 orders of magnitude smaller than the thickness of Unit 4-W (Table 1). The modeled thickness of dry gouge varies from 4.2 μm at a slip velocity of 1.3 m/s to 50 μm at 0.1 m/s. These values are 85% and 40% of the measured thicknesses of Unit 4-D at $v_r = 0.7$ (5 μm) and 1.3 m/s (15 μm), respectively. At a slip velocity of 0.35 m/s, the modeled thickness of 12 μm is much smaller than the overall thickness of Unit 4-D but similar to the thickness of an individual lamina. We propose that active localized slip actually only occurs within a micrometer scale zone, defined here as a lamina, and that this zone of slip is progressively abandoned as slip migrates to a new zone, leaving behind a sequence of stacked, inactive micrometer-scale shear zones.

In addition to thermal weakening by frictional heating, slip weakening and localization can occur in granular media by grain size reduction (Beeler et al., 1996; Hadizadeh et al., 2015). In clay-rich rocks, the localized slip zone thickness that develops at subseismic rates is between 10 and 100 times the clay platelet size (Morgenstern & Tchalenko, 1967). We take the clay platelet size of Unit 3, which is $\sim 0.1 \mu\text{m}$, as an upper bound on the median platelet size of Unit 4-W. The localized slip zone is, therefore, expected to be a maximum of 1 to 10 μm thick, which is 1% to 10% the thickness of Unit 4-W. In Unit 4-D, the average size of measurable clay grains is similar to Unit 3. However, there is a significant fraction of clay material with a grain size that is too small to resolve. Given the resolution of the TEM images, this material has a maximum median grain size of $\sim 0.01 \mu\text{m}$ if it is crystalline, resulting in a predicted slip zone thickness of 1 to 10 μm , which spans the lamina thickness of Unit 4-D (Figure 7) and is also similar to the thickness predicted by thermal dehydration weakening.

We conclude that the thickness of Unit 4-W is most consistent with modeled slip zone thicknesses that develop during thermal pressurization. The differences between modeled and measured thicknesses may

reflect a combination of gradient in slip velocity in, poorly controlled boundary conditions, as pore fluids are not completely sealed, and thermal conduction through the dolerite host blocks. The apparently slip- and rate-weakening mechanical properties of wet gouge are both consistent with thermal pressurization (French et al., 2014). Thermal pressurization of water-wet gouge is also consistent with the temperature rise in the gouge layer, which did not reach the dehydration temperature in most experiments but did reach water vaporization temperatures at 1-MPa normal stress (Figure 2b). It is possible that some thermal dehydration accompanied shear of Unit 4-W, because the temperatures at 0.7 and 1.3 m/s approached the dehydration temperature of smectite and may have been locally higher than calculated values which assume distributed deformation.

At the normal stress conditions of the experiments, heating of pore water can lead to vaporization which would then be expected to result in gouge strengthening if all pore water is vaporized. However, such behavior is not observed in these and other experiments where vaporization conditions are reached (Boutareaud et al., 2010; Kitajima et al., 2010). In French et al. (2014) we used the thermomechanical models to estimate the amount of mechanical energy that was dissipated as sensible heat versus latent heat by processes that include fracture, pore fluid vaporization, and dehydration (Table 1). The energy that would be consumed by vaporization of all added pore water is $E_{vp} = m_w \times \Delta H_v$, where m_w is the mass of pore water (0.3 g) and ΔH_v is the latent heat of vaporization (2,260 kJ/mol). Thus, $E_{vp} = 678$ J, which is more than all latent heat dissipated at 0.7 and 1.3 m/s and 93% of that dissipated at 0.35 m/s. It follows that not all pore water was vaporized in the experiments but that a significant fraction of the latent heat may have been consumed by vaporization.

The thickness and structure of Unit 4-D is consistent with both thermal dehydration and granular shear localization if the active localized slip zone migrates during deformation, and we infer that it develops as a combination of the two processes. The fact that the only shear rate that did not result in localization is 0.1 m/s, and this is also the only rate at which the temperature rise was clearly insufficient for thermal dehydration suggests a causal relationship between temperature rise, dehydration, and localization (Figure 2d). We suggest that migration occurs when the actively shearing zone is sufficiently dehydrated and/or grain size is sufficiently reduced that shearing migrates to a new weaker layer. Unfortunately, as a result of changes in hydration state that can occur during sample storage, during section preparation, and as a result of electron beam damage, it is near impossible to determine the experimental hydration state of the saponite using TEM images and diffraction patterns. It is not obvious why Unit 4-D contains many more laminae in the sample sheared at 0.35 m/s (#2420). The total energy dissipated by this sample is the greatest (Table 1), although not proportional to differences in total thickness of Unit 4-D. One possibility is that the energy partitioning between grain size reduction and dehydration varies with deformation rate and controls migration of deformation. As done for water-wet gouge, we investigate the amount of energy potentially dissipated by thermal dehydration of Unit 4-D and compare it to our previous estimates of latent heat dissipation to evaluate whether this is a feasible mechanism of weakening and localization.

The amount of latent heat dissipated by dehydration is $E_{dhy} = v \times f_{sm} \times \rho / m_{sm} \times E_a$, where v is the volume of the dehydrating zone, $f_{sm} = 0.9$ is the mass fraction of smectite, ρ is smectite density, m_{sm} is smectite molar mass, and E_a is the activation energy of dehydration (Brantut et al., 2008). Although activation energy for saponite dehydration is not available, those of montmorillonite dehydration reactions vary from ~36 and 50 kJ/mol (Bray & Redfern, 1999; Van Groos & Guggenheim, 1986). The density and molar mass of smectite are 2,080 kg/m³ and 0.549 kg/mol, respectively. We estimate the maximum energy consumed by dehydration of Unit 4-D by calculating the total volume of Unit 4-D in each sample. The total energy ranges from 0.2 J at $v_r = 1.3$ m/s to 15 J at $v_r = 0.35$ m/s, corresponding to 0.02% and 1.3% of the latent heat we previously determined, respectively (Table 1). The maximum energy that can be consumed by dehydration of the entire CDZ layer is 130 J, which is still only 12–13% of the latent heat in experiments that produce a Unit 4-D. We draw two conclusions based on these calculations. The first is that our hypothesis that shear migrates when the reactant is depleted is feasible given that more than enough energy is consumed for the dehydration reactions. The second is that a significant amount of energy is likely also consumed by cataclastic or other mechanical degradation process given that chemical processes can only account for a small fraction of the latent heat.

5.3. Fabric and Gouge Kinematics

The foliation orientations differ between the deformation units reflecting evolving and heterogeneous strain within the entire deforming gouge layer. The P and Y foliations defined by clay shear surfaces in Unit 2 are

commonly reported in naturally and experimentally sheared clay gouge (Figure 3b; Chester & Logan, 1987; Logan et al., 1979; Rutter et al., 1986). Our previous results show that this composite foliation develops in Unit 2 during subseismic slip rates and seismic slip rates over small shear strains (French et al., 2014).

5.3.1. Kinematics of Localized Slip: Unit 4

We analyze the orientations of lamina in Unit 4-D to understand the kinematics of localized slip (Figure 9). We infer, based on clast and matrix particle size reduction in Unit 4-D relative to Unit 3 (Figure 5), that the gouge within the lamina has been sheared and therefore slip did not simply occur along the lamina. Deformation of laminae in an R orientation alone, however, cannot accommodate significant shear strain and also is accommodated by extension and thinning of Unit 4-D during shear. The second set of lamina in a P orientation addresses these strain compatibility issues and supports our hypothesis that they develop as a system to maintain near-constant gouge layer thickness under conditions of very high simple shear strain. We have fewer microstructural constraints on whether Unit 4-W deformed by homogeneous shear or as migrating, stacking slip zones. However, given the lack of sharp boundaries that indicate discontinuities in strain, we favor the hypothesis that Unit 4-W deformed homogeneously.

5.3.2. Kinematics of Fluidization: Unit 3

The isotropic fabric of the fluidized gouge in Unit 3 and the lack of observable fragmentation counterparts are consistent with descriptions of fluidized fault-rock textures (Boutareaud et al., 2010; Monzawa & Otsuki, 2003; Otsuki et al., 2003; Rowe et al., 2005; Ujiie et al., 2011). To estimate the shear strain rate accommodated by Unit 3, we employ empirical relations for the viscosity of a fluidized gouge, ν_g , and then calculate the shear strain rate as $\dot{\epsilon} = \tau / \nu_g$ (Bagnold, 1954; Monzawa & Otsuki, 2003). Powder technology research shows that viscosity of fluidized solid grains is a function of the volume fraction of grains, ϕ , and this work has been applied to fault gouges (Bagnold, 1954; Monzawa & Otsuki, 2003). An empirical relation for ν_g is

$$\nu_g = \nu_f \left(1 - \frac{\phi}{\phi^*} \right)^{-K\phi^*} \quad (3)$$

where ν_f is the viscosity of the interstitial fluid, ϕ is the volume fraction of grains, ϕ^* is the maximum possible volume fraction of grains, and K is an empirical constant. Monzawa and Otsuki (2003) noted that the exponent $-K\phi^*$ is frequently found to be ~ 2 .

Strain was dilatant during shearing at 0.35 to 1.3 m/s representative velocity, indicating that ϕ decreases with slip to a near constant value (Figures 2b and 2d). At the end of each experiment, the entire gouge layer compacted, presumably during dissipation of excess pore fluid pressure and pore collapse (Table 1). The orientations of clasts within Unit 3 are consistent with a flattening fabric, but the orientations of matrix particles suggest that they did not rotate significantly during compaction (Figures 4e–4g). Though the lack of a compaction fabric in the clay is perplexing, Meneghini et al. (2010) showed that platy minerals in a naturally deformed black fault rock inferred by Rowe et al. (2005) to be fluidized have no preferred orientation. If shortening at the end of each experiment, δw , reflects a transition in grain volume fraction from ϕ during shear to ϕ^* , then $\phi/\phi^* = 1 - \delta w/1.2$, where 1.2 mm is the thickness of the gouge layer. We assume that dilation during shear occurs uniformly throughout the gouge layer, giving a lower bound on the dilation of Unit 3, which we predict dilates more than Unit 4 (Table 1). For water vapor viscosity of 1.5×10^{-5} Pa s, equation (3) predicts that the viscosity of Unit 3 varies from 0.005 to 0.05 Pa s in these experiments. Taking the steady state shear strengths from Figure 2, the shear strain rates in the fluidized gouge zones range from 5 to 45 s⁻¹, which is only 0.6 to 10% of the strain rates accommodated by each entire gouge layer and therefore the remainder occurs within Unit 4.

Our conclusion that the majority of shear strain occurs within Unit 4 means that the highest shear heating rate and thermally produced pore fluid pressure occurs within Unit 4 as well. A gradient in fluid pressure between Units 4 and 3 will generate a resulting flow of pore fluid. This flow is consistent with evidence of injection structures that originate at the boundary with Unit 4 and lead to the development of Unit 3 (French et al., 2014). The concentration of matrix particles that are oriented normal to the shear plane in Unit 3 is consistent with development under fluidized conditions where elevated pore fluid pressure is generated within Unit 4 (Figure 4c). Thus, although evidence for localized slip is not sufficient to infer coseismic slip, the paired occurrence of a localized shear zone and a zone distinguished by an isotropic fabric can provide strong support for a history of seismic slip. This conclusion is supported because the isotropic fabric provides evidence of suspension and transport of particles associated with fluid movement through the granular gouge that in these experiments requires coseismic shear heating and associated pore fluid pressurization.

6. Lithologic Controls of Slip Localization

Evidence for localized slip is common in both the products of experimentally sheared gouges (e.g., Mizoguchi et al., 2007; Ujiie et al., 2011) and exhumed fault zones (e.g., Chester & Chester, 1998; Reed, 1964; Sibson, 1977). Although localization is not itself sufficient to infer seismic slip, the architecture of localized slip zones and co-occurrence of structural features within these zones can provide evidence for a history of earthquake slip that may be useful in the absence of unequivocal evidence like frictional melting. In clay-rich fault rock, records of frictional heating during seismic slip are challenging to preserve unless the rock melted, in part because dehydration is a reversible process and decomposition products can be difficult to identify as such. Recent detailed structural analyses of carbonate fault gouges and rocks have shown that dynamic weakening processes can be recorded in the fabric of the fault rock (Bullock et al., 2015; Rempe et al., 2014; Smith et al., 2017), and herein we present additional evidence that the structures that form in clay-rich fault rocks at seismic slip rates and low effective normal stress are distinct from those that have been shown to develop at subseismic deformation rates (e.g., French et al., 2015; Logan et al., 1979).

The character of microfoliated zones that develop during experiments on carbonate (Bullock et al., 2015; Boulton et al., 2017; Rempe et al., 2014; Smith et al., 2017), exhumed clay-bearing (Kitajima et al., 2010; Mizoguchi et al., 2007) and granitic (Stunitz et al., 2010) gouges are not as sensitive to pore fluid content as the CDZ gouge and other clay-rich gouges, and the fabric of most of these zones resemble that described for our Unit 4-D. Similar to our Unit 4-D, the localized slip zones in these other lithologies are distinguished by a fine grain size and laminae that have been interpreted to show internal variations in porosity (Rempe et al., 2017; Stunitz et al., 2010). Laminated slip zones in the ultracataclasite of the Punchbowl fault have also been interpreted to develop by progressive wear and accumulation (Chester et al., 2004), which is consistent with our observations that the inferred older laminae show increasing density and decreasing porosity. In carbonate and clay gouges, very fine-grained localized slip zones have been interpreted to form as a result of extreme particle size reduction and additional chemical decomposition (e.g., Brantut et al., 2008; Paola et al., 2015; Han et al., 2007; Green et al., 2015; Mitchell et al., 2015). Our calculations of localized slip zone thickness and energy dissipation indicate that cataclastic deformation plays a significant, if not dominant, role in mechanical energy dissipation of clay-rich gouges and slip zone thickness. The relative importance of cataclasis over decomposition in reducing grain size and forming laminae is illustrated by the development of similar laminae at subseismic slip rates but large displacements (Hadizadeh et al., 2015; Scruggs & Tullis, 1998). Therefore, this structure alone is questionable evidence of seismic slip unless a record of thermal decomposition is preserved (Brantut & Sulem, 2012; Mitchell et al., 2015; Smith et al., 2017).

Clay-rich gouges, including the CDZ gouge, that were deformed mainly at subseismic rates tend to develop a distributed foliation defined by R, Y, and P shears, but not the stacked laminae that form in Unit 4-D of our clay-rich gouge, and in other lithologies over a wide range of deformation conditions and saturation states (French et al., 2015; Logan et al., 1979; Rutter et al., 1986). Thus, any record of laminae composed of very fine-grained clay-rich fault gouge may indeed provide evidence of seismicity. More structural observations of clay-rich gouge sheared at subseismic rates and high displacement would clarify the uniqueness of this feature to seismic slip rates. The development of the foliation seen in Unit 4-W seems to be limited to fault gouges that are composed predominately of clay minerals and sheared with a pore water (e.g., Boulton et al., 2017; Boutareaud et al., 2010; Ujiie & Tsutsumi, 2010). Interpreting this foliation as having developed during seismic slip may be challenging, however, because it resembles fabrics also developed at subseismic slip rates but here occurs over a narrow zone. If a similar fabric develops in a natural fault zone, it may be difficult to distinguish this zone from the surrounding foliated rock unless it is paired with a fluidized zone, like our Unit 3. At the low normal stresses of these experiments, it is likely that the formation of Unit 3 was enhanced by vaporization of water, so that it may not develop at depths below a few kilometers where vaporization is suppressed by pressure except where low normal stress occurs locally due to nonplanar fault geometry. We conclude that the pairing of localized slip with a fluidized gouge layer provides evidence in favor of seismic slip in any lithology. This evidence may be useful in interpreting the history of fault slip, particularly when frictional melt is absent, and other potential structural records of seismic slip are present (Rowe & Griffith, 2015). We note that Kirkpatrick et al. (2015); Keren and Kirkpatrick (2016) identified localized slip zones that are qualitatively similar to our Unit 4-W in fault core recovered from the megathrust that hosted the 2011 Tokoku earthquake. The same core also contains regions lacking a clear shear fabric or particle shape preferred orientation that may be similar to our Unit 3 (Kirkpatrick et al., 2015).

6.1. Conclusions

We measured localized slip zone thicknesses, particle orientations, and particle size distributions in clay-rich fault gouge from the CDZ deformed at high slip rates in the laboratory to (1) determine how heterogeneous strain is partitioned in seismically shearing fault gouge, (2) test existing numerical models for the thickness and character of the localized slip zones that develop at seismic rates, and (3) relate the microscale structures that develop at seismic rates to the processes that formed them.

1. Comparison of model calculations with measured thicknesses of localized slip zones show that the slip zone thickness in Unit 4-W is consistent with predictions for thermal pressurization in wet gouge. In dry gouge, the thickness of individual laminae within the localized slip zone (Unit 4-D) are consistent with both thermal dehydration and cataclasis. The total thickness of the slip zone is concluded to increase during slip as actively shearing laminae are abandoned and new ones form. Grain sizes and energy calculations for dry gouge also indicate that cataclasis plays an important role in deformation of these localized, high shear-strain slip zones.
2. Clast orientations and estimates of fluidized layer viscosity in Unit 3 indicate that it accommodates at most 10% of the total shear strain and usually an order of magnitude less. Therefore, the fluidized gouge accumulates very little of the coseismic shear strain, most of which accumulates within Unit 4.
3. Measurements of particle size distributions show that clast sizes become more uniform with increasing shear strain. Within the laminae of Unit 4-D, clasts larger than the grinding limit are more uniform than has been reported for seismically sheared rocks (PSD exponents up to 6). Clasts smaller than the grinding limit have a less uniform size distribution (PSD exponents ~ 1). We interpret this pattern to reflect spalling of the small fragments off larger clasts. Clay particles, in contrast, show no clear trend in PSD with interpreted strain of the unit.
4. The pairing of a localized high shear-strain zone with a random-fabric gouge zone supports the interpretation of past seismic slip in clay-rich fault gouge. This pairing may be unique to the low normal stresses of these experiments, as higher normal stresses suppress the vaporization that contribute to fluidization. In the absence of a fluidized gouge layer, the microfoliations composed of μ -scale lamina in Unit 4-D may be unique to seismic slip rates, as they have not been reported for clay-rich gouge deformed at subseismic rates.

Appendix A

A1. Model Calculations of Thermal Pressurization

Rice et al. (2014) presented a linearized analytic model for the thickness of the principal slipping zone during thermal pressurization, and Platt et al. (2014) presented a numerical simulation of the same problem and found a more precise but similar result. We present the calculated thickness of localized shear zones in wet CDZ gouge using results from Platt et al. (2014), but we note that we found the calculated thickness to be within 7% using the formulation of Rice et al. (2014) at seismic strain rates. The model parameter values used and the relevant references are provided in Table A1. The results of the models are in Table 1. The localized shear zone thickness determined by Platt et al. (2014) is

$$W \approx \frac{6.9(a-b)\rho c}{f_0 + 2(a-b)} \frac{(\sqrt{\alpha_{hy}} + \sqrt{\alpha_{th}})^2}{Vf_0\Lambda} \quad (A1)$$

where $a-b$ is the steady state friction rate dependence and f_0 is the friction coefficient. ρ is density, c is specific heat, V is the slip rate, α_{hy} is hydraulic diffusivity, and α_{th} is thermal diffusivity. Λ is a measure of the change in fluid pressure with a change in temperature and is given by $\Lambda = (\lambda_f - \lambda_n)/(\beta_f + \beta_n)$, where λ is compressibility, β is thermal expansivity, and the subscripts f and n refers to fluid and pore space, respectively. We calculate $\alpha_{th} = k/(v_f\beta)$, where k is the permeability, v_f is the fluid viscosity, and $\beta = n(\beta_f + \beta_n)$ is the storage coefficient.

A2. Model Calculations of Thermal Decomposition Pressurization

We parameterized the model for localization during thermal decomposition (Platt et al., 2015) with both wet and dry gouge properties. The localized slip zone thickness is

$$W = 0.55(p + [q + (q^2 - p^6)^{1/2}]^{1/3} + [q - (q^2 - p^6)^{1/2}]^{1/3}) \quad (A2)$$

Table A1*Parameters Used for Models of Thermal Pressurization (TP) and Decomposition Weakening (DW) and References for the Parameters Used*

Saturation		TP wet, Platt et al. (2014)	DW wet, Platt et al. (2015)	DW dry, Platt et al. (2015)	Reference
$a - b$		0.003	0.003	0.005	Carpenter et al. (2012), Lockner et al. (2011)
f_0		0.2	0.2	0.3	French et al. (2014) (0.1 m/s)
n	Calculated from experiments of French et al. (2014)				
V	mm/s	350, 700, 1,300			French et al. (2014)
ρ	kg/m ³	2,300	2,300	2,300	
ρ_f	kg/m ³	800	800	800	
c	J/(kg·K)	1,000	1,000	1,000	Michot et al. (2008)
α_{th}	mm ² /s	0.7	0.7	0.7	Rice et al. (2014)
Λ	MPa/K	0.3	0.3	0.3	Rice et al. (2014)
$\lambda_f (\times 10^{-3})$	C ⁻¹	3.1	3.1	3.1	Platt et al. (2014)
$\lambda_n (\times 10^{-3})$	C ⁻¹	-1.4	-1.4	-1.4	Platt et al. (2014)
$\beta_f (\times 10^{-9})$	Pa ⁻¹	0.64	0.64	0.64	Rice (2006)
$\beta_n (\times 10^{-9})$	Pa ⁻¹	2.5	2.5	2.5	Rice (2006)
k	m ²	10 ⁻¹⁸	10 ⁻¹⁸	10 ⁻¹⁸	Morrow et al. (2014)
A	1/s		10 ¹⁰	10 ¹⁰	Bray and Redfern (1999)
Q	kJ/mol		46	46	Van Groos and Guggenheim (1986)
ΔH	MJ/kg		3.3	3.3	Van Groos and Guggenheim (1986)
m_d	kg/m ³		77	77	Bray and Redfern (1999)
ϕ	m ³ /kg		5 × 10 ⁻⁴	5 × 10 ⁻⁴	
T_f	K		493	493	see text

where

$$p = \frac{\pi \alpha_{hy} E_r (a - b) \rho c}{3f_0^2 V (P_r - \Lambda E_r)}, q = \frac{\pi^4 \alpha_{hy} \alpha_{th} (a - b) \rho c}{2f_0^2 V \bar{m} (P_r - \Lambda E_r) \beta_1 \beta_2} \quad (A3)$$

β_1 and β_2 characterize the reaction kinetics and are $\beta_1 = A \exp -\frac{Q}{RT_f}$ and $\beta_2 = \exp \frac{Q}{RT_f}$, respectively, where A is the preexponential factor, Q is activation energy, T_f is the temperature at which the reaction initiates, and R is the gas constant. For the smectites, dehydration occurs through the progressive loss of interlayer water over a temperature range with two peaks in water loss, the first occurring with the greatest water loss (Van Groos & Guggenheim, 1984). The temperatures of these peaks varies according to the smectite member, cations present, and pressure. Although there is not sufficient data for saponite at 1 MPa pressure, data consistently show that the first dehydration peak occurs between 100 and 140 °C at atmospheric pressure and increases with pressure to approximately 200 to 240 °C at 1 MPa (Bray & Redfern, 1999; Van Groos & Guggenheim, 1984, 1986; Kawano & Tomita, 1991). Thus, we use $T_f = 493$ K for our calculations. Measurements of Q for different montmorillonite compositions vary between 36 and 46 kJ/mol, and unfortunately we know of no published values for saponite (Van Groos & Guggenheim, 1984, 1986). However, we find that even doubling this activation energy which more than accounts for the more thermally stable behavior of saponite results in a less than 1% change in our result. The parameter E_r is the temperature change of dehydration of a pure smectite gouge and is given by $\Delta H m_d / (\rho c)$, where ΔH is the enthalpy of the reaction and m_d is the mass of pore fluid released during dehydration of pure smectite. P_r is the pore pressure produced by the reaction when boundary conditions are undrained and adiabatic and is equal to $1/(\rho_f \phi)(1 - \rho_f \phi) m_d$, where ϕ is the pore volume increase per mass of water released during dehydration. $\bar{m} = m_{dg}/m_d$ and m_{dg} is the mass of water released for complete reaction per unit volume of gouge.

Acknowledgments

This research was supported by the NSF EarthScope program (EAR-0643339) and the Southern California Earthquake Center (Contribution 8882). SCEC is funded by NSF Cooperative Agreement EAR-1033462 and USGS Cooperative Agreement G12AC20038. We want to thank the Japan Agency for Marine-Earth Science and Technology for offering us an opportunity to work in their laboratory and Caleb Holyoke for assistance with the TEM. We thank Diane Moore and an anonymous reviewer for comments that helped to improve the manuscript. The data required to reproduce the results are provided in Tables 1 and 2 or available for download from the Rice Digital Scholarship Archive (<https://doi.org/10.25611/kytg-mw69>).

References

- Abe, S., & Mair, K. (2005). Grain fracture in 3D numerical simulations of granular shear. *Geophysical Research Letters*, 32, L05305. <https://doi.org/10.1029/2004GL022123>
- Agterberg, F. P. (1974). *Geomathematics. Mathematical background and geo-science applications* (596 pp.). Amsterdam, New York: Elsevier Scientific Pub. Co.
- An, L.-J., & Sammis, C. G. (1994). Particle size distribution of cataclastic fault materials from Southern California: A 3-D study. *Pure and Applied Geophysics*, 143(1), 203–227. <https://doi.org/10.1007/BF00874329>
- Aretusini, S., Mitterpergher, S., Plumper, O., Spagnuolo, E., Gualtieri, A., & Toro, G. D. (2017). Production of nanoparticles during experimental deformation of smectite and implications for seismic slip. *Earth and Planetary Science Letters*, 463, 221–231. <https://doi.org/10.1016/j.epsl.2017.01.048>
- Bagnold, R. (1954). Experiments on a gravity-free dispersion of large solid spheres in a Newtonian fluid under shear, Proceedings of the Royal Society of London A: Mathematical, Physical and Engineering Sciences, 225(1160), 49–63. <https://doi.org/10.1098/rspa.1954.0186>
- Beeler, N. M., Tullis, T. E., Blanpied, M. L., & Weeks, J. D. (1996). Frictional behavior of large displacement experimental faults. *Journal of Geophysical Research*, 101(B4), 8697–8715.
- Ben-Zion, Y., & Sammis, C. G. (2003). Characterization of fault zones. *Pure and Applied Geophysics*, 160(3), 677–715. <https://doi.org/10.1007/PL00012554>
- Boulton, C., Yao, L., Faulkner, D. R., Townend, J., Toy, V. G., Sutherland, R., et al. (2017). High-velocity frictional properties of alpine fault rocks: Mechanical data, microstructural analysis, and implications for rupture propagation. *Journal of Structural Geology*, 97, 71–92. <https://doi.org/10.1016/j.jsg.2017.02.003>
- Boutareaud, S., Boullier, A.-M., Andreani, M., Calugaru, D.-G., Beck, P., Song, S.-R., & Shimamoto, T. (2010). Clay clast aggregates in gouges: New textural evidence for seismic faulting. *Journal of Geophysical Research*, 115, B02408. <https://doi.org/10.1029/2008JB006254>
- Boutareaud, S., Hirose, T., Andreani, M., Pec, M., Calugaru, D.-G., Boullier, A.-M., & Doan, M.-L. (2012). On the role of phyllosilicates on fault lubrication: Insight from micro- and nanostructural investigations on talc friction experiments. *Journal of Geophysical Research*, 117, B08408. <https://doi.org/10.1029/2011JB009006>
- Bradbury, K. K., Evans, J. P., Chester, J. S., Chester, F. M., & Kirschner, D. L. (2011). Lithology and internal structure of the San Andreas fault at depth based on characterization of phase 3 whole-rock core in the San Andreas Fault Observatory at Depth (SAFOD) borehole. *Earth and Planetary Science Letters*, 310(1–2), 131–144. <https://doi.org/10.1016/j.epsl.2011.07.020>
- Brantut, N., Schubnel, A., Rouzaud, J. N., Brunet, F., & Shimamoto, T. (2008). High-velocity frictional properties of a clay-bearing fault gouge and implications for earthquake mechanics. *Journal of Geophysical Research*, 113, B10401. <https://doi.org/10.1029/2007JB005551>
- Brantut, N., & Sulem, J. (2012). Strain localization and slip instability in a strain-rate hardening, chemically weakening material. *Journal of Applied Mechanics*, 79(3), 10. <https://doi.org/10.1115/1.4005880>
- Bray, H. J., & Redfern, S. A. T. (1999). Kinetics of dehydration of Ca-montmorillonite. *Physics and Chemistry of Minerals*, 26(7), 591–600. <https://doi.org/10.1007/s002690050223>
- Bullock, R. J., Paola, N. D., & Holdsworth, R. E. (2015). An experimental investigation into the role of phyllosilicate content on earthquake propagation during seismic slip in carbonate faults. *Journal of Geophysical Research: Solid Earth*, 120, 3187–3207. <https://doi.org/10.1002/2015JB011914>
- Carpenter, B. M., Saffer, D. M., & Marone, C. (2012). Frictional properties and sliding stability of the San Andreas fault from deep drill core. *Geology*, 40, 759–762. <https://doi.org/10.1130/G33007.1>
- Chester, F. M., & Chester, J. S. (1998). Ultracataclastic structure and friction processes of the Punchbowl Fault, San Andreas system, California. *Tectonophysics*, 295(1–2), 199–221.
- Chester, F., Chester, J., Kirschner, D., Schulz, S., & Evans, J. (2004). Structure of large-displacement, strike-slip fault zones in the brittle continental crust. In *Rheology and Deformation in the Lithosphere at Continental Margins*, 1 (pp. 223–260).
- Chester, J. S., Chester, F. M., & Kronenberg, A. K. (2005). Fracture surface energy of the Punchbowl Fault, San Andreas system. *Nature*, 437(7055), 133–136. <https://doi.org/10.1038/nature03942>
- Chester, F. M., & Logan, J. M. (1986). Implications for mechanical properties of brittle faults from observations of the Punchbowl Fault zone, California. *Pure and Applied Geophysics*, 124(1–2), 79–106.
- Chester, F. M., & Logan, J. M. (1987). Composite planar fabric of gouge from the Punchbowl Fault, California. *Journal of Structural Geology*, 9(5–6), 621–634.
- Cladouhos, T. T. (1999). Shape preferred orientations of survivor grains in fault gouge. *Journal of Structural Geology*, 21(4), 419–436. [https://doi.org/10.1016/S0191-8141\(98\)00123-0](https://doi.org/10.1016/S0191-8141(98)00123-0)
- Coble, C. G., French, M. E., Chester, F. M., Chester, J. S., & Kitajima, H. (2014). In situ frictional properties of San Andreas fault gouge at SAFOD. *Geophysical Journal International*, 199(2), 956–967. <https://doi.org/10.1093/gji/ggu306>
- Cowan, D. S. (1999). Do faults preserve a record of seismic slip? A field geologist's opinion. *Journal of Structural Geology*, 21(8–9), 995–1001.
- Di Toro, G., & Pennacchioni, G. (2005). Fault plane processes and mesoscopic structure of a strong-type seismogenic fault in tonalites (Adamello batholith, Southern Alps). *Tectonophysics*, 402(1), 55–80. <https://doi.org/10.1016/j.tecto.2004.12.036>
- Dieterich, J. (1978). Time-dependent friction and the mechanics of stick-slip. *Pure and Applied Geophysics*, 116(4–5), 790–806.
- Faulkner, D., Lewis, A., & Rutter, E. (2003). On the internal structure and mechanics of large strike-slip fault zones: Field observations of the Carboneras fault in southeastern Spain. *Tectonophysics*, 367(3), 235–251. [https://doi.org/10.1016/S0040-1951\(03\)00134-3](https://doi.org/10.1016/S0040-1951(03)00134-3)
- French, M. E., Chester, F. M., & Chester, J. S. (2015). Micromechanisms of creep in clay-rich gouge from the Central Deforming Zone of the San Andreas fault. *Journal of Geophysical Research: Solid Earth*, 120, 827–849. <https://doi.org/10.1002/2014JB011496>
- French, M. E., Kitajima, H., Chester, J. S., Chester, F. M., & Hirose, T. (2014). Displacement and dynamic weakening processes in smectite-rich gouge from the Central Deforming Zone of the San Andreas fault. *Journal of Geophysical Research: Solid Earth*, 119, 1777–1802. <https://doi.org/10.1002/2013JB010757>
- Ghosh, S., & Ramberg, H. (1976). Reorientation of inclusions by combination of pure shear and simple shear. *Tectonophysics*, 34(1), 1–70. [https://doi.org/10.1016/0040-1951\(76\)90176-1](https://doi.org/10.1016/0040-1951(76)90176-1)
- Green, H. W. II, Shi, F., Bozhilov, K., Xia, G., & Reches, Z. (2015). Phase transformation and nanometric flow cause extreme weakening during fault slip. *Nature Geoscience*, 8(6), 484–489.
- Hadizadeh, J., & Johnson, W. K. (2003). Estimating local strain due to comminution in experimental cataclastic textures. *Journal of Structural Geology*, 25(11), 1973–1979. [https://doi.org/10.1016/S0191-8141\(03\)00016-6](https://doi.org/10.1016/S0191-8141(03)00016-6)
- Hadizadeh, J., Tullis, T. E., White, J. C., & Konkachbaev, A. I. (2015). Shear localization, velocity weakening behavior, and development of cataclastic foliation in experimental granite gouge. *Journal of Structural Geology*, 71, 86–99. <https://doi.org/10.1016/j.jsg.2014.10.013>
- Han, R., Shimamoto, T., Hirose, T., Ree, J.-H., & Ando, J.-i. (2007). Ultralow friction of carbonate faults caused by thermal decomposition. *Science*, 316(5826), 878–881. <https://doi.org/10.1126/science.1139763>

- Heilbronner, R., & Keulen, N. (2006). Grain size and grain shape analysis of fault rocks. *Tectonophysics*, 427(1-4), 199–216. <https://doi.org/10.1016/j.tecto.2006.05.020>
- Hirose, T., Mizoguchi, K., & Shimamoto, T. (2012). Wear processes in rocks at slow to high slip rates. *Journal of Structural Geology*, 38, 102–116. <https://doi.org/10.1016/j.jsg.2011.12.007>
- Ikari, M. J., Carpenter, B. M., Vogt, C., & Kopf, A. J. (2016). Elevated time-dependent strengthening rates observed in San Andreas fault drilling samples. *Earth and Planetary Science Letters*, 450, 164–172. <https://doi.org/10.1016/j.epsl.2016.06.036>
- Kawano, M., & Tomita, K. (1991). Dehydration and rehydration of saponite and vermiculite. *Clays and Clay Minerals*, 39(2), 174–183.
- Keren, T. T., & Kirkpatrick, J. D. (2016). The damage is done: Low fault friction recorded in the damage zone of the shallow Japan Trench décollement. *Journal of Geophysical Research: Solid Earth*, 121, 3804–3824. <https://doi.org/10.1002/2015JB012311>
- Keulen, N., Heilbronner, R., Stunitz, H., Boullier, A.-M., & Ito, H. (2007). Grain size distributions of fault rocks: A comparison between experimentally and naturally deformed granitoids. *Journal of Structural Geology*, 29(8), 1282–1300. <https://doi.org/10.1016/j.jsg.2007.04.003>
- Kirkpatrick, J. D., Rowe, C. D., Ujiie, K., Moore, J. C., Regalla, C., Remitti, F., et al. (2015). Structure and lithology of the Japan Trench subduction plate boundary fault. *Tectonics*, 34, 53–69. <https://doi.org/10.1002/2014TC003695>
- Kitajima, H., Chester, J. S., Chester, F. M., & Shimamoto, T. (2010). High-speed friction of disaggregated ultracataclasite in rotary shear: Characterization of frictional heating, mechanical behavior, and microstructure evolution. *Journal of Geophysical Research*, 115, B08408. <https://doi.org/10.1029/2009JB007038>
- Lockner, D. A., Morrow, C., Moore, D., & Hickman, S. (2011). Low strength of deep San Andreas fault gouge from SAFOD core. *Nature*, 472(7341), 82–85. <https://doi.org/10.1038/nature09927>
- Logan, J. M., Friedman, M., Higgs, N., Dengo, C., & Shimamoto, T. (1979). Experimental studies of simulated gouge and their application to studies of natural fault zones. Open-File Report - U.S. Geological Survey, pp. 305–343.
- Marone, C., & Scholz, C. (1989). Particle-size distribution and microstructures within simulated fault gouge. *Journal of Structural Geology*, 11(7), 799–814. [https://doi.org/10.1016/0191-8141\(89\)90099-0](https://doi.org/10.1016/0191-8141(89)90099-0)
- Meneghini, F., Di Toro, G., Rowe, C., Moore, J., Tsutsumi, A., & Yamaguchi, A. (2010). Record of mega-earthquakes in subduction thrusts: The black fault rocks of Pasagshak Point (Kodiak Island, Alaska). *Geological Society of America Bulletin*, 122(7-8), 1280–1297. <https://doi.org/10.1130/B30049.1>
- Michot, A., Smith, D. S., Degot, S., & Gault, C. (2008). Thermal conductivity and specific heat of kaolinite: Evolution with thermal treatment. *Journal of the European Ceramic Society*, 28(14), 2639–2644. <https://doi.org/10.1016/j.jeurceramsoc.2008.04.007>
- Mitchell, T. M., Smith, S. A., Anders, M. H., Toro, G. D., Nielsen, S., Cavallo, A., & Beard, A. D. (2015). Catastrophic emplacement of giant landslides aided by thermal decomposition: Heart Mountain, Wyoming. *Earth and Planetary Science Letters*, 411, 199–207. <https://doi.org/10.1016/j.epsl.2014.10.051>
- Mizoguchi, K., Hirose, T., Shimamoto, T., & Fukuyama, E. (2007). Reconstruction of seismic faulting by high-velocity friction experiments: An example of the 1995 Kobe earthquake. *Geophysical Research Letters*, 34, L01308. <https://doi.org/10.1029/2006GL027931>
- Mizoguchi, K., Hirose, T., Shimamoto, T., & Fukuyama, E. (2009). High-velocity frictional behavior and microstructure evolution of fault gouge obtained from Nojima fault, southwest Japan. *Tectonophysics*, 471(3-4), 285–296. <https://doi.org/10.1016/j.tecto.2009.02.033>
- Monzawa, N., & Otsuki, K. (2003). Commintion and fluidization of granular fault materials: Implications for fault slip behavior. *Tectonophysics*, 367(1-2), 127–143. [https://doi.org/10.1016/S0040-1951\(03\)00133-](https://doi.org/10.1016/S0040-1951(03)00133-)
- Moore, D. E. (2014). Comparative mineral chemistry and textures of SAFOD fault gouge and damage-zone rocks. *Journal of Structural Geology*, 68, 82–96. <https://doi.org/10.1016/j.jsg.2014.09.002>
- Moore, J. C., & Byrne, T. (1987). Thickening of fault zones: A mechanism of melange formation in accreting sediments. *Geology*, 15(11), 1040–1043. [https://doi.org/10.1130/0091-7613\(1987\)15<1040:TOFZAM>2.0.CO;2](https://doi.org/10.1130/0091-7613(1987)15<1040:TOFZAM>2.0.CO;2)
- Morgenstern, N., & Tchalenko, J. (1967). Microscopic structures in kaolin subjected to direct shear. *Géotechnique*, 17(4), 309–328. <https://doi.org/10.1680/geot.1967.17.4.309>
- Morrow, C., Lockner, D., Moore, D., & Hickman, S. (2014). Deep permeability of the San Andreas fault from San Andreas Fault Observatory at Depth (SAFOD) core samples. *Journal of Structural Geology*, 64, 99–114. <https://doi.org/10.1016/j.jsg.2013.09.009>
- Mühlhaus, H. B., & Vardoulakis, I. (1987). The thickness of shear bands in granular materials. *Géotechnique*, 37(3), 271–283. <https://doi.org/10.1680/geot.1987.37.3.271>
- Otsuki, K., Monzawa, N., & Nagase, T. (2003). Fluidization and melting of fault gouge during seismic slip: Identification in the Nojima fault zone and implications for focal earthquake mechanisms. *Journal of Geophysical Research*, 108(B4), 2192. <https://doi.org/10.1029/2001JB001711>
- Paola, N. D., Holdsworth, R. E., Viti, C., Colletini, C., & Bullock, R. (2015). Can grain size sensitive flow lubricate faults during the initial stages of earthquake propagation? *Earth and Planetary Science Letters*, 431, 48–58. <https://doi.org/10.1016/j.epsl.2015.09.002>
- Platt, J. D., Brantut, N., & Rice, J. R. (2015). Strain localization driven by thermal decomposition during seismic shear. *Journal of Geophysical Research: Solid Earth*, 120, 4405–4433. <https://doi.org/10.1002/2014JB011493>
- Platt, J. D., Rudnicki, J. W., & Rice, J. R. (2014). Stability and localization of rapid shear in fluid-saturated fault gouge: 2. Localized zone width and strength evolution. *Journal of Geophysical Research: Solid Earth*, 119, 4334–4359. <https://doi.org/10.1002/2013JB010711>
- Prasher, C. (1987). *Crushing and grinding process handbook*. Chichester, UK: Wiley.
- Reed, J. J. (1964). Mylonites, cataclasites, and associated rocks along the Alpine Fault, South Island, New Zealand. *New Zealand Journal of Geology and Geophysics*, 7(4), 645–684. <https://doi.org/10.1080/00288306.1964.10428124>
- Rempe, M., Smith, S. A., Ferri, F., Mitchell, T. M., & Toro, G. D. (2014). Clast-cortex aggregates in experimental and natural calcite-bearing fault zones. *Journal of Structural Geology*, 68, 142–157. <https://doi.org/10.1016/j.jsg.2014.09.007>
- Rempe, M., Smith, S., Mitchell, T., Hirose, T., & Toro, G. D. (2017). The effect of water on strain localization in calcite fault gouge sheared at seismic slip rates. *Journal of Structural Geology*, 97, 104–117. <https://doi.org/10.1016/j.jsg.2017.02.007>
- Rice, J. R. (2006). Heating and weakening of faults during earthquake slip. *Journal of Geophysical Research*, 111, B05311. <https://doi.org/10.1029/2005JB004006>
- Rice, J. R., Rudnicki, J. W., & Platt, J. D. (2014). Stability and localization of rapid shear in fluid-saturated fault gouge: 1. Linearized stability analysis. *Journal of Geophysical Research: Solid Earth*, 119, 4311–4333. <https://doi.org/10.1002/2013JB010710>
- Rowe, C. D., & Griffith, W. A. (2015). Do faults preserve a record of seismic slip: A second opinion. *Journal of Structural Geology*, 78, 1–26. <https://doi.org/10.1016/j.jsg.2015.06.006>
- Rowe, C. D., Meneghini, F., & Moore, J. C. (2011). Textural record of the seismic cycle: Strain-rate variation in an ancient subduction thrust. *Geological Society*, 359(1), 77–95. <https://doi.org/10.1144/SP359.5>
- Rowe, C. D., Moore, J. C., Meneghini, F., & McKeirnan, A. W. (2005). Large-scale pseudotachylites and fluidized cataclasites from an ancient subduction thrust fault. *Geology*, 33(12), 937–940. <https://doi.org/10.1130/G21856.1>

- Rowe, C. D., Moore, J. C., & Remitti, F. A. (2013). The thickness of subduction plate boundary faults from the seafloor into the seismogenic zone. *Geology*, 41(9), 991–994. <https://doi.org/10.1130/G34556.1>
- Rutter, E. H., Maddock, R. H., Hall, S. H., & White, S. H. (1986). Comparative microstructures of natural and experimentally produced clay-bearing fault gouges. *Pure and Applied Geophysics*, 124(1-2), 3–30.
- Sammis, C., King, G., & Biegel, R. (1987). The kinematics of gouge deformation. *Pure and Applied Geophysics*, 125(5), 777–812. <https://doi.org/10.1007/BF00878033>
- Schleicher, A. M., van der Pluijm, B. A., & Warr, L. N. (2010). Nanocoatings of clay and creep of the San Andreas fault at Parkfield, California. *Geology*, 38(7), 667–670. <https://doi.org/10.1130/G31091.1>
- Schleicher, A. M., van der Pluijm, B. A., & Warr, L. N. (2012). Chlorite-smectite clay minerals and fault behavior: New evidence from the San Andreas Fault Observatory at Depth (SAFOD) core. *Lithosphere*, 4, 209–220. <https://doi.org/10.1130/L158.1>
- Schleicher, A. M., Warr, L. N., & van der Pluijm, B. A. (2009). On the origin of mixed layered clay minerals from the San Andreas fault at 2.5–3 km vertical depth (SAFOD drillhole at Parkfield, California). *Contributions to Mineralogy and Petrology*, 157(2), 173–187. <https://doi.org/10.1007/s00410-008-0328-7>
- Scruggs, V., & Tullis, T. (1998). Correlation between velocity dependence of friction and strain localization in large displacement experiments on feldspar, muscovite and biotite gouge. *Tectonophysics*, 295(1), 15–40. [https://doi.org/10.1016/S0040-1951\(98\)00113-9](https://doi.org/10.1016/S0040-1951(98)00113-9)
- Shimamoto, T., & Tsutsumi, A. (1994). A new rotary-shear high-speed frictional testing machine: Its basic design and scope of research. *Journal of Tectonic Research Group of Japan*, 39, 69–78.
- Sibson, R. H. (1977). Fault rocks and fault mechanisms. *Journal of the Geological Society*, 133(3), 191–213. <https://doi.org/10.1144/gsjgs.133.3.0191>
- Sibson, R. H. (2003). Thickness of the seismic slip zone. *Bulletin of the Seismological Society of America*, 93(3), 1169–1178. <https://doi.org/10.1785/0120020061>
- Smith, S. A. F., Billi, A., DiToro, G., & Spiess, R. (2011). Principal slip zones in limestone: Microstructural characterization and implications for the seismic cycle (Tre Monti fault, central Apennines, Italy). *Pure and Applied Geophysics*, 168(12), 2365–2393. <https://doi.org/10.1007/s00024-011-0267-5>
- Smith, S. A., Griffiths, J. R., Fondriest, M., & DiToro, G. (2017). 'Coseismic foliations' in gouge and cataclasis: Experimental observations and consequences for interpreting the fault rock record. In M. Y. Thomas (Ed.), *Fault dynamic processes evolution of fault properties during seismic rupture* (pp. 81–102). Hoboken, NJ: John Wiley. <https://doi.org/10.1002/9781119156895>
- Storti, F., Billi, A., & Salvini, F. (2003). Particle size distributions in natural carbonate fault rocks: Insights for non-self-similar cataclasis. *Earth and Planetary Science Letters*, 206(1), 173–186. [https://doi.org/10.1016/S0012-821X\(02\)01077-4](https://doi.org/10.1016/S0012-821X(02)01077-4)
- Stunitz, H., Keulen, N., Hirose, T., & Heilbronner, R. (2010). Grain size distribution and microstructures of experimentally sheared granitoid gouge at coseismic slip rates; criteria to distinguish seismic and aseismic faults? *Journal of Structural Geology*, 32(1), 59–69.
- Turcotte, D. (1997). *Fractals and chaos in geology and geophysics*. New York: Cambridge University Press.
- Ujiie, K., & Tsutsumi, A. (2010). High-velocity frictional properties of clay-rich fault gouge in a megaseismic fault zone, Nankai subduction zone. *Geophysical Research Letters*, 37, L24310. <https://doi.org/10.1029/2010GL046002>
- Ujiie, K., Tsutsumi, A., & Kameda, J. (2011). Reproduction of thermal pressurization and fluidization of clay-rich fault gouges by high-velocity friction experiments and implications for seismic slip in natural faults. *Geological Society Special Publications*, 359, 267–285. <https://doi.org/10.1144/SP359.15>
- Ujiie, K., Yamaguchi, A., Kimura, G., & Toh, S. (2007). Fluidization of granular material in a subduction thrust at seismogenic depths. *Earth and Planetary Science Letters*, 259(3), 307–318. <https://doi.org/10.1016/j.epsl.2007.04.049>
- Van Groos, A. K., & Guggenheim, S. (1984). The effect of pressure on the dehydration reaction of interlayer water in Na-montmorillonite (SWY-1). *American Mineralogist*, 69(9-10), 872–879.
- Van Groos, A. K., & Guggenheim, S. (1986). Dehydration of K-exchanged montmorillonite at elevated temperatures and pressures. *Clays and Clay Minerals*, 34(3), 281–286.
- Wibberley, C. A. J., & Shimamoto, T. (2005). Earthquake slip weakening and asperities explained by thermal pressurization. *Nature*, 436(7051), 689–692. <https://doi.org/10.1038/nature03901>
- Wilson, B., Dewers, T., Reches, Z., & Brune, J. (2005). Particle size and energetics of gouge from earthquake rupture zones. *Nature*, 434, 749–752. <https://doi.org/10.1038/nature03433>
- Yao, L., Ma, S., Shimamoto, T., & Togo, T. (2013). Structures and high-velocity frictional properties of the Pingxi fault zone in the Longmenshan fault system, Sichuan, China, activated during the 2008 Wenchuan earthquake. *Tectonophysics*, 599, 135–156. <https://doi.org/10.1016/j.tecto.2013.04.011>