

High-velocity frictional properties of a clay-bearing fault gouge and implications for earthquake mechanics

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[1] Frictional properties of natural kaolinite-bearing gouge samples from the Median Tectonic Line (SW Japan) have been studied using a high-velocity rotary shear apparatus, and deformed samples have been observed with optical and electron (scanning and transmission) microscopy. For a slip velocity of 1 m s⁻¹ and normal stresses from 0.3 to 1.3 MPa, a dramatic slip-weakening behavior was observed. X-ray diffraction analysis of deformed samples and additional high-velocity friction experiments on pure kaolinite indicate kaolinite dehydration during slip. The critical slip-weakening distance D_c is of the order of 1 to 10 m. These values are extrapolated to higher normal stresses, assuming that D_c is rather a thermal parameter than a parameter related to a true characteristic length. The calculation shows that dimensionally, $D_c \propto 1/\sigma_n^2$, where σ_n is the normal stress applied on the fault. The inferred D_c values range from a few centimeters at 10 MPa normal stress to a few hundreds of microns at 100 MPa normal stress. Microscopic observations show partial amorphization and dramatic grain size reduction (down to the nanometer scale) localized in a narrow zone of about 1 to 10 μ m thickness. Fracture energy G_c is calculated from the mechanical curves and compared to surface energy due to grain size reduction, and energies of mineralogic transformations. We show that most of the fracture energy is either converted into heat or radiated energy. The geophysical consequences of thermal dehydration of bonded water during seismic slip are then commented in the light of mineralogical and poromechanical data of several fault zones, which tend to show that this phenomenon has to be taken into account in most of subsurface faults and in hydrous rocks of subducted oceanic crust.

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1. Introduction

[2] Rocks occurring within shallow faults usually display an incohesive gouge zone, which corresponds to the wear product of slip between fault surfaces. These fault gouges often contain an important amount of various clays (e.g., recent results from San Andreas Fault Observatory at Depth (SAFOD) [*Solum et al.*, 2006]). The mechanical properties of these rocks play a crucial role in the source mechanism of earthquakes. In particular, frictional properties of fault gouges are related to the evolution of coseismic slip, its instability and the energies involved in slip.

[3] However, since the 1960s, most of the experimental work on rock friction has been conducted at low slip velocities (from 1 μ m s⁻¹ to several centimeters per second), which has led to the rate-and-state (R-S) variable

friction laws [Dietrich, 1978; Ruina, 1983]. Although R-S laws are widely used to understand and model seismic instability, it is likely that they may provide an inadequate framework because coseismic slip occurs at velocities of the order of 1 m s⁻¹ and thermal effects become extremely relevant under these deformation conditions [e.g., Rice 2006]. This inconsistency is particularly emphasized by the discrepancy on the actual value for the critical slip parameter D_c , which corresponds to the slip distance after which the frictional strength of a fault is dramatically reduced. D_c values inferred from seismological data for natural earthquakes are of the order of 1 to 10 m [Ide and Takeo, 1997; Mikumo et al., 2003; Fukuyama et al., 2003], whereas laboratory measurements at low slip velocities lead to smaller values by orders of magnitude, from 10^{-5} to 10^{-3} [Dietrich, 1978, 1979; Ohnaka, 1992; Marone and Kilgore, 1993]. Several explanations have been proposed to solve this paradox. Morphological parameters such as variations in fault surface roughness [Scholz, 1988] and thickness of the deformed zone [Marone and Kilgore, 1993], mechanical properties such as the size of the nucleation patch, total slip distance [Ohnaka, 2003] and dynamically induced off-fault yielding [Andrews, 1976, 2005] and thermal effects such as

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Figure 1. Geological setting of Tsukide outcrop on the MTL. (a) Location of the MTL outcrop in Tsukide. (b) Photograph of the outcrop (R stands for the Ryoke unit, S for the Sambagawa unit); the dashed line highlights the central slip zone; the scale is 2 m.

flash heating on fault asperities [*Rice*, 2006] are plausible explanations for this discrepancy.

[4] In the same way, the argument over the correlation between the total breakdown work during an earthquake, as observed by seismologists, and the fracture surface energy observed on the field, has been vivid [Tinti et al., 2005]. For example, field work performed on the Punchbowl fault indicates that the total surface energy, including the damage and process zones surrounding the fault, may well be greater than the observed fracture energy in the case of a single seismic rupture [Chester et al., 2005], while Ma et al. [2006] have recently shown that the surface energy linked to grain fragmentation and comminution in fault cores from the Chelungpu fault drilling represented only 6% of the total fracture energy observed during the Chi-Chi earthquake. In a more general way, the energy balance during an earthquake, and the processes involved in this budget, remains an open question, and much experimental work is needed to understand the micromechanical processes taking place during an earthquake.

[5] Recent experiments that have been conducted with a high-velocity frictional rotary shear apparatus have led to D_c measurements of a few meters in saw cut rock material [Tsutsumi and Shimamoto, 1997; Hirose and Shimamoto, 2003, 2005], i.e., much closer to seismically inferred values. Recent works using the same apparatus show similar results on natural fault gouges coming from Nojima fault [Mizoguchi et al., 2007]. These friction experiments employing high slip velocities have all emphasized the significance of frictional heating and possible thermohydromechanical coupling during high-velocity friction (HVF) tests via various dynamic weakening mechanisms: for example, partial melting (pseudotachylytes) for dry crystalline rocks [e.g., Hirose and Shimamoto, 2005; Di Toro et al., 2006], decarbonation of marble [Han et al., 2007b] or siderite [Han et al., 2007a], or silica gel production in quartz gouges [Goldsby and Tullis, 2002; Di Toro et al., 2004]. These experimental observations show the large diversity in deformation mechanisms and the physical processes involved in friction of rocks at seismic slip rates.

[6] Of particular importance, it seems that experimental evidence for frictional exsolution of fluids (water, CO₂) during coseismic slip may well be partially confirmed by recent field evidence of carbonate degassing in Nojima fault [Famin et al., 2008] and clay minerals changes in gouge samples from Chelungpu borehole [Song et al., 2007]. This leads to important consequences, as fluid release might provide additional source terms in the thermal pressurization (TP) equations. Indeed, TP of pore fluids generally leads to a poroelastic increase of pore pressure due to frictional heating during slip, as initially suggested by Sibson [1973]. Whereas TP is particularly important in low-permeability rocks such as clay-bearing fault gouges and has significant implications both for the value of D_c [Wibberley and Shimamoto, 2005; Rice, 2006] and for the energy budget of rupture itself [Rempel and Rice, 2006; Rice, 2006], friction-induced release of fluids during coseismic slip may be nonnegligible in gouges containing hydrous minerals.

[7] All these observations emphasize the complexity and variety of friction related phenomena in fault rocks, and the importance of thermohydromechanical coupling in these processes. In light of these recent observations, this study presents experimental results on the mechanical and chemical behavior of a natural fault gouge (from the Median Tectonic Line, MTL, SW Japan) during HVF, and discusses their theoretical implications for the overall energy budget. The consequences of coseismic dehydration of hydrous minerals such as clays are then discussed by comparing the pore pressure rise due to TP to the one due to possible mineral dehydration.

2. Experimental Procedure

2.1. Sample Collection and Preparation

[8] The Median Tectonic Line is a major strike-slip fault, which intersects two metamorphic belts: in the north, the Ryoke Belt (low pressure, high temperature) and in the south, the Sambagawa Belt (high pressure, low temperature). The MTL is still active in its western part (in Shikoku and the western Kii peninsula). Its motion, which started in



Figure 2. Exposure and mineralogy of the fault gouges. (a) Detailed view of the central slip zone. (b) Interpreted sketch (c, Ryoke-derived coarse white gouge; g1 to g3, Sambagawa-derived gouge layers); scale bar is 2 cm. (c) Raw X-ray diffraction patterns of gouges 1 (external) and gouge 3 (internal); k stands for kaolinite peaks, q stands for quartz peaks, d stands for dolomite, and s stands for siderite. Note the absence of kaolinite and siderite in the central gouge layer (g3).

the early Cretaceous, is quite complex and beyond the scope of this work [*Ichikawa*, 1980].

[9] The outcrop studied here is located in Tsukide (Mie Prefecture) on the eastern part of the fault (see Figure 1a). Information about the internal structure, the microtexture and some poromechanical parameters of the fault rocks have already been published by Suwa et al. [1997] and Wibberley and Shimamoto [2003], and we will here briefly summarize these observations. Figure 1b shows a very clear contact zone between two different terrains. Near the central slip zone, the Sambagawa metapelitic rocks form several gouge layers a few centimeters in thickness (Figures 2a and 2b). The external gouge (g1) is foliated in the same orientation displayed by the fault plane, and contains numerous clasts of various sizes, from several millimeters to several centimeters in diameter. The internal gouge (g3) contains no visible clasts. X-ray diffraction (XRD) measurements (CuK α radiation, with an accelerating voltage of 40 kV and a filament current of 20 mA) were performed on powdered gouge samples coming from this outcrop. The diffraction patterns (Figure 2c) reveal the presence of quartz, dolomite, siderite, kaolinite and albite in the external gouge (g1). The kaolinite mass fraction was inferred independently from mineral mode calculation on the basis of bulk chemical analysis performed with energy dispersive X-ray

spectrometry (EDS) on a scanning electron microscope (SEM), and by comparing the XRD pattern of the gouge to XRD patterns of known mixtures of kaolinite and quartz. The diffractometer used for these quantitative measurements is a Rigaku UltraX 18 (rotating anode), with beam conditions set at 40 kV and 300 mA. This leads to a mass fraction of kaolinite of $\sim 10-20\%$. In the internal gouge (g3), kaolinite is absent, and there is a strong decrease in dolomite and siderite content compared to the external gouge. These observations may suggest that the g1 gouge could be the parent material of the g3 gouge, which may have been formed during a seismic event. Following this hypothesis, g1 gouge was the one that was deformed during the seismic event, and therefore this gouge layer was chosen for our HVF experiments.

[10] Gouge samples (g1) were collected and tested with a rotary shear, high-speed frictional apparatus [*Shimamoto and Tsutsumi*, 1994]. For each experiment, one gram of the gouge was first powdered and dried at ambient conditions, at room temperature during a few days, and then placed between two host rock cylinders (gabbro, \sim 25 mm in diameter, with end surfaces ground with SiC 80 powder). In addition, a few pure kaolinite artificial gouge samples were also prepared. In each experiment, a Teflon ring was



Figure 3. High-velocity rotary shear apparatus. (a) Simplified sketch of the high-velocity frictional testing machine: 1, sample; 2, motor; 3, torque limiter; 4, torque gauge; 5, electromagnetic clutch; 6, rotary encoder; 7, rotary column; 8, torque-axial force gauge; 9, spline; 10, axial force gauge; 11, air actuator; 12, displacement transducer; 13, moisture sensor. Modified from *Hirose and Shimamoto* [2005] and *Mizoguchi* [2005]. (b) Detailed sketch of the sample assembly.

placed around the gouge to confine it during shearing (Figure 3b).

2.2. Experimental Setup and Method

[11] The experimental device is shown on Figure 3a (modified from Hirose and Shimamoto [2005] and Mizoguchi [2005]). Once the sample was prepared, the cylindrical sample assembly (gouge powder plus gabbro plus Teflon ring) is placed at position (1) within the loading column. The normal load is applied with an air actuator (11). Only the left part of the sample can rotate at high velocity (up to 1500 rpm) thanks to a 7.5 kW motor (2); the right part of the sample is fixed to a spline (9) that allows for horizontal movement only. An electromagnetic clutch (5) is used to couple the rotation of the motor to the load column; this device allows us to start rotating the sample immediately at the desired speed. The rotation speed is measured by a rotary encoder (6). The torque is measured with torque gauges (4, 8): only the one on the right, stationary side is used for torque measurement, since the left one (moving side) is influenced by friction on the loading column. Axial load is measured with a force gauge at position (10). Finally, axial shortening is measured with a displacement transducer at position (12). In addition, for some experiments we used a moisture sensor (13) to record the evolution of humidity in the vicinity of the sample; the distance between sample and sensor was of about 3 cm. This sensor was used only for the experiments performed on pure kaolinite, in order to check the evolution of water vapor content near the sample. For such experiments, the sample chamber was sealed by installing a transparent plastic window over the frame with tape and clay.

[12] All the data (rotation speed, axial load, torque, axial shortening and relative humidity) was recorded with an analog chart recorder and a digital recorder (TEAC, DR-f1) at a rate of 200 Hz. Numerical data were filtered with a voltage-controlled voltage-source (VCVS)-type noise filter with a 50-Hz high-frequency cutoff.

[13] Because of the sample geometry, the slip rate varied as a function of distance from the center of the rotation axis. In the following, an equivalent slip velocity v_{eq} is defined such that $\tau v_{eq}S$ is the rate of frictional work on the slipping

surface area S, assuming that the shear stress τ is constant over the entire fault surface [Shimamoto and Tsutsumi, 1994]. This velocity is given by

$$v_{\rm eq} = \frac{4\pi}{3}\Omega R$$

where *R* is the radius of the sample and Ω the angular velocity. Experiments on the MTL gouge samples were performed at an equivalent velocity of 1.04 m s⁻¹ (i.e., at an angular velocity of 1200 rpm), normal stresses of 0.28 to 1.32 MPa and for total displacements of 3 to 54 m. A summary of the experimental conditions is given for each experiment in Table 1. In most of these experiments, a first run without applying any normal stress was performed to make the Teflon ring smoother and looser and thus reduce the friction associated with it. It is assumed that this step did not change the overall behavior of the sample in subsequent tests.

3. Results

3.1. Mechanical Data

[14] A series of 13 HVF tests were conducted on samples of the MTL gouge layer (g1) at different normal stresses, ranging from 0.28 to 1.32 MPa and at v_{eq} of 1.04 m s⁻¹. The typical mechanical behavior of the MTL gouge samples is represented in Figure 4. Figure 4 displays the apparent frictional coefficient μ (ratio of shear stress to normal stress) as a function of total displacement.

[15] In our experiments, a dramatic decrease in shear strength was observed. This slip-weakening behavior can be described using only three parameters: the apparent frictional coefficient at peak friction, $\mu_{\rm p}$, the residual friction coefficient $\mu_{\rm r}$ and the critical slip weakening distance D_c . Using these parameters, the data can be fitted using the following empirical equation [*Hirose and Shimamoto*, 2005; *Mizoguchi et al.*, 2007]:

$$\mu_{\rm obs} = \mu_r + \left(\mu_{\rm p} - \mu_{\rm r}\right) e^{\ln(0.05)d/D_c} \tag{1}$$

	Equivalent Velocity (m s ⁻¹)	Normal Stress (MPa)	Total Displacement (m)	μ Peak	μ Residual	D_c (m)
HVR784 ^b	1.04	0.28	54.1	1.23	0.50	50.2
HVR787 ^b	1.04	0.48	39.5	1.09	0.40	18.0
HVR782 ^b	1.04	0.59	48.1	1.02	0.28	16.6
HVR786 ^b	1.04	0.82	37.3	0.89	0.23	20.5
HVR781 ^b	1.04	0.99	37.5	0.77	0.24	19.0
HVR788 ^b	1.04	1.15	31.2	0.82	0.21	13.1
HVR780 ^b	1.04	1.32	36.2	0.85	0.26	4.1
HVR754	1.04	0.60	43.7	1.29	0.34	12.5
HVR861	1.04	0.61	3.0	1.06	-	-
HVR863	1.04	0.61	4.5	1.23	-	-
HVR865	1.04	0.60	7.4	1.30	-	-
HVR866	1.04	0.61	7.8	1.02	-	-
HVR905	1.04	0.62	38.1	1.30	0.27	6.3

Table 1. Experimental Conditions^a

^aExperiments performed on samples from the MTL (g1) fault gouge.

^bExperiments for which the Teflon friction was reduced (see text for details).

 $\mu_{\rm T}$ and D_c are determined by the fitting method and $\mu_{\rm p}$ corresponds to the value observed at the peak stress. The parameter D_c is defined here as the slip distance needed to achieve 95% of the total weakening. This definition makes this parameter comparable to the D_c parameter inferred from seismological measurements, which represents the distance for the fault to weaken during an earthquake. For each experiment, the values observed for $\mu_{\rm p}$ and D_c are reported as a function of normal stress in Table 1. In our experiments, the apparent friction at peak $\mu_{\rm p}$ ranged from 0.77 to 1.29, the residual friction $\mu_{\rm r}$ from 0.21 to 0.5 and the critical slip weakening distance D_c from 4 to 50 m. It is important to note that these raw data are not corrected for the Teflon friction before fitting.

[16] We observe a linear dependence between shear stress and normal stress, which is consistent with a Mohr-Coulomb type friction law (Figure 5a):

$$\tau = \mu \sigma_n + C \tag{2}$$

where τ is the shear stress, σ_n the normal stress, μ the "intrinsic" friction coefficient and *C* the cohesive term. A linear fit of our data gives the following values:

$$\tau_{\rm p} = 0.69\sigma_n + 0.17\tag{3}$$

$$\tau_{\rm r} = 0.15\sigma_n + 0.10 \tag{4}$$

where τ_p is the peak shear stress and τ_r is the residual shear stress. Here our experimental results highlight the fact that the "intrinsic" friction coefficient is dramatically reduced from ~0.7 to ~0.15 during the HVF experiments. The cohesive terms are relatively small in value and since our samples are initially noncohesive, they can be attributed mainly to the presence of the Teflon ring [*Mizoguchi et al.*, 2007], which does not support any axial load (see Figure 3b) or to weak Van der Waals or hydrogen bonds between grains. Figure 5b shows that the critical slip weakening distance D_c also decreases from ~50 m down to ~4 m with increasing normal stress (up to 1.36 MPa). These observations are consistent with similar data previously obtained on Nojima fault gouge samples by *Mizoguchi et al.* [2007].

3.2. Mineralogical and Chemical Data

[17] After the experiments, the gouge samples were extracted, disaggregated and characterized using XRD. XRD pattern measurements were performed on two different gouge samples, sheared respectively for 7.8 m (HVR866) and 43.7 m (HVR754) of displacement (at an equivalent velocity of 1.04 m s⁻¹ and a normal stress of 0.60 MPa). Given that the outer edge of the gouge samples underwent the largest amount of shear, only gouge products from this outer part (a few millimeters in width) were analyzed. Because gouge extraction was performed manually, the analyzed gouge powder contained some of the less or nondeformed gouge. Figure 6 displays the measured diffraction peaks, normalized to the main quartz peak. The peak for quartz is considered as a reference, because we



Figure 4. Typical evolution of the friction coefficient of an MTL fault gouge sample (experiment HVR754) during a HVF experiment. The curve displays a peak friction coefficient μ_p , which drops to a residual friction coefficient μ_r once the critical weakening distance D_c is attained. The dashed line corresponds to best data fit by the empirical equation (1). Also displayed are G_c the fracture energy and *H* the "heating energy"; both are discussed in section 4.



Figure 5. Evolutions of (a) peak and residual shear stress and (b) D_c as a function of normal stress. Figure 5a shows linear dependences between the peak and the residual shear stresses and the normal stress. The linear fit outputs the values of the "intrinsic" frictional coefficients (at peak stress and residual) and the cohesion term. The latter is mainly due to the presence of the Teflon ring. Figure 5b shows the critical slip weakening distance D_c decreases to a few meters with increasing normal stress. The dashed line is a fit following the dependency derived in equation (8).

observed that most of the quartz grains were still intact within the extracted gouge. Figure 6 highlights that the intensities of the two main kaolinite peaks drastically decrease with increasing total displacement. In fact, for sample HVR754 (43.7 m total displacement) the kaolinite peaks have almost completely disappeared. This might indicate that kaolinite minerals are being decomposed during shearing.

[18] HVF experiments were also performed on artificial gouge samples of pure kaolinite. During these experiments, relative humidity was measured in the sample chamber (see Figure 3). For these experiments, the sample chamber was sealed. Its watertightness was checked by performing runs without any sample in the device, and no change of relative humidity was then observed. Figure 7 displays representative evolution of friction, relative humidity and axial displacement with increasing displacement for a pure kaolinite gouge sample tested at $v_{eq} = 1.3 \text{ m s}^{-1}$ and $\sigma_n = 0.63 \text{ MPa}$ (HVR808). The overall frictional behavior of pure kaolinite gouge samples is similar to that of the MTL gouge. The corresponding axial displacement decreases at the beginning of the run as the sample shortens, and finally increases starting from 35 m slip, indicating dilatancy. Relative humidity increases of more than 2% during the experiment indicate that the kaolinite gouge sample lost water during shearing. The final dilatancy of the gouge could be related to this water release, indicating that there might be a small pore pressure increase that could counteract the axial shortening of the sample. The water that is released during the experiment may have two distinct origins: (1) adsorbed water, between kaolinite grains and (2) structural water, bonded as OH radicals. Kaolinite is a 1:1 (one tetrahedral layer bonded to one octahedral layer) layer silicate, with very few ionic substitution, thus the amount of interlayer water is not significant [Deer et al., 1966; Giese, 1988]. This along with the X-ray diffraction

patterns suggest that at least part of the released water comes from the thermal decomposition of kaolinite, following the reaction

$$Al_2Si_2O_5(OH)_4 \rightarrow \underbrace{Al_2Si_2O_7}_{metakaolin} + 2H_2O$$
 (5)

Under dry conditions (argon atmosphere), the dehydration temperature of kaolinite ranges from 480 to 590°C [*Frost et al.*, 2004; *Horváth et al.*, 2003]. Besides water, the



Figure 6. Mineralogical evolution of the MTL fault gouge sample with increasing displacement. The X-ray diffraction patterns of two deformed samples are displayed; the peak intensities are normalized to the main quartz peak. Intensities of the two main kaolinite peaks decrease clearly with increasing displacement.



Figure 7. Evolution of relative humidity, axial shortening, and friction with slip for pure kaolinite gouge samples. The black curve shows an increase of relative humidity (water vapor content) during the HVF experiment. The frictional behavior (gray curve) of pure kaolinite is similar to that of the MTL fault gouge. During the experiment, the relative humidity increases more than 2% in the sample chamber. The axial displacement (red curve) first decreases as the sample shortens and finally increases after ~35 m slip.

decomposition product is disordered metakaolin, an amorphous phase which does not appear on X-ray diffraction patterns. The possibility of kaolinite dehydration during high-velocity shearing will be further investigated in section 4.2.2.

3.3. Microstructural Analysis

[19] Thin sections cut through deformed gouge samples after HVF experiments (oriented as shown on Figure 3b) were observed using various microscopy techniques: optical microscopy, scanning electron microscopy (SEM) and transmission electron microscopy (TEM).

[20] On the optical micrographs (Figure 8a) a thin, darker zone located along the host rock can be distinguished, which varies in thickness from a couple to tens of microns. Given that there are no other apparent deformation markers inside the gouge, this zone is interpreted as the main slipping zone [Mizoguchi, 2005]. Within this zone, the grain size is below SEM resolution (0.1 μ m)(Figures 8b and 8c). Thus, in order to investigate the internal structure of the slipping zone, a focused ion beam (FIB) section was extracted from the gouge. The FIB apparatus uses a focused beam of gallium ions to cut a part of the sample and thin it in order to produce a thin section of about one hundred nanometers in thickness, which can then be observed under the TEM. The deformed gouge sample analyzed in the following was sheared at an equivalent velocity of 1.03 m s^{-1} , a normal stress of 1.43 MPa, and for a total displacement of 37 m (HVR780). The FIB section was cut perpendicularly to the slipping zone as shown on Figure 9a.

[21] TEM micrographs (Figure 9b) reveal that the slipping zone contains small grains (≤ 100 nm) only, embedded within a gray matrix which appears morphologically homogeneous at the TEM scale. Note that the initial gouge matrix contained grains larger than a few hundreds of microns. This confirms the dramatic grain size reduction (about 3 orders of magnitude) that took place within the slipping zone during HVF tests. In order to specify the important mineralogical and crystallographic changes due to deformation, elemental composition mapping was first performed using the scanning transmission electron microscopy (STEM) mode of the TEM to obtain a local chemical information on the remaining minerals and their matrix within the slipping zone (Figure 10); the probe diameter is about 20 nm, corresponding to analyzed volumes of about $10^{-3} \ \mu m^3$. Then selected area electron diffraction (SAED) patterns were recorded on volumes of about $10^{-2} \ \mu m^3$ in order to distinguish crystallized and amorphous constituents of the slipping zone.

[22] On the TEM micrographs, distinct grains, chemically and geometrically well defined, can be observed within the



Figure 8. Microtexture of an MTL gouge sample after HVF test. The sample was sheared at $v_{eq} = 1.04 \text{ m s}^{-1}$ and $\sigma_n = 1.34 \text{ MPa}$, for total displacement of 37 m. (a) Microphotograph of the artificial gouge after shearing, showing a thin and highly deformed zone considered to be the main slipping zone. (b) SEM image highlighting the smooth, compacted surface of the slipping zone. (c) Detailed SEM image. Note the apparent absence of distinct grains. The presence of glue is an artifact due to the thin section preparation.



Figure 9. FIB section cutting and TEM observation. (a) After the cutting, the section is soldered to a platinum needle, extracted and soldered again to a TEM copper grid: 1, platinum needle; 2, relatively undeformed gouge; 3, glue; 4, slipping zone; 5, FIB section; and 6, host rock. (b) TEM picture (same orientation): small grains (less than 100 nm) can be distinguished inside a gray matrix which seems homogeneous at this scale. Frames 1' and 2' were analyzed respectively with X-ray mapping and electron diffraction techniques.

slipping zone (Figure 10): (1) numerous small grains $(\sim 10 \text{ nm in size})$ containing titanium (possibly rutile) and (2) larger grains (~ 100 nm) containing iron and magnesium, often associated with grains also containing calcium. Since chemical mapping did not reveal the presence of silicon or carbon atoms within these grains, they are probably metal oxides. The matrix however, is mainly composed of silicon and aluminum with some zones containing silicon exclusively; such elemental analyses suggest that these areas can be derived from quartz, feldspar and clay minerals present in the initial gouge sample. At this scale, the texture of the matrix again appears quite homogeneous, and it is important to point out the absence of both kaolinite or well defined quartz grains within the slipping zone. Thus, these observations are consistent with the X-ray diffraction patterns presented previously.

[23] In order to further investigate the evolution of the crystalline state of the materials that compose the slipping zone, TEM images and SAED patterns were acquired. Figure 11 presents a TEM micrograph taken at the bottom of the slipping zone (frame 2' on Figure 9b). Here, the microtexture is quite heterogeneous: some grains display Bragg fringes (formed by superimposed and slightly misoriented crystallites), which indicates that these crystals are only slightly deformed. Electron diffraction patterns from outside the grains display broad and faint diffusion rings (patterns 1, 2, and 3), and some diffraction spots (patterns 2) and 3). The diffusion ring highlights the presence of amorphous materials within the slipping zone; the ring position depends on the average interatomic distance. The diffraction spots indicate the presence of submicron-sized crystals which cannot be distinguished on the TEM image, probably due to superposition of crystals across the FIB section thickness. Inside the grains, electron diffraction patterns (patterns 4 and 5) display well-organized diffraction spots thus demonstrating that they are crystalline. Note once again that no trace of crystallized kaolinite was found anywhere within the FIB section. Similar TEM investigations were also performed on powdered nondeformed gouge

samples and revealed that the initial gouge contained only well defined, well crystallized grains (mainly quartz grains), and some intact kaolinite minerals.

[24] Our microstructural observations prove the presence of a thin (~1 to 10 μ m), ultralocalized deformation zone where grain size was dramatically reduced to the submicron scale. We interpret this zone as being the main slipping zone active during HVF. The occurrence of the amorphous matrix in the gouge slipping zone is probably the result of a pronounced amorphization of the quartz and kaolinite crystals. This is suggested by the TEM data from the FIB section cut from this zone: disappearance of the (hkl) reflections in the SAED patterns coupled with the Si-Al composition as shown by STEM mapping. The hardness of the titanium oxides is probably responsible for the observed relics of Ti-based nanocrystals. Importantly, our observations are consistent with previous microstructural analysis performed on experimental fault gouges in granite, quartzite and marble [Yund et al., 1990] deformed at low slip speeds. After high shear strains, our experiments also show that the deformed gouge contained a significant amount of amorphous material and submicron-sized grains. The amorphized part of the slipping zone does not contain a significant amount of sodium or potassium, which would normally be present in the case of partial melting of feldspar. Moreover, our observations do not indicate any textural evidence of partial melting of the gouge. Thus, we suggest that amorphization might be related to extreme grain comminution [Yund et al., 1990].

4. Interpretations and Discussion

4.1. Scaling Laws and Extrapolating D_c to High Normal Stresses

[25] *Ohnaka* [2003] suggested the following scaling law to correlate mechanical and geometric parameters of a fault:

$$\frac{\Delta \tau_{\rm b}}{\tau_{\rm p}} = \beta \left(\frac{D_c}{\lambda_c}\right)^M \tag{6}$$



Figure 10. X-ray mapping in the slipping zone. The top left image is the reference picture obtained under the TEM, in scanning electron mode (STEM). Each consecutive image reveals the relative atomic density of each of the following elements: Si, Al, Ca, Mg, Fe, Ti, and K, respectively. Different types of grains can be distinguished: Ti concentrates within small grains a few tens of nanometers in size, whereas Fe, Mg, and Ca are associated with larger grains hundreds of nanometers in size. The bottom right image is an interpreted sketch: grains that contain Ti are represented in black, those that contain Fe and Mg are represented in dark gray, and those that contain Ca are represented in light gray. Parts that contain Si only are highlighted by the dashed lines.

where $\Delta \tau_b = \tau_p - \tau_r$ is the breakdown stress drop, λ_c is a characteristic length scale of the fault surface, and β and M are constants. Using the values calculated by Ohnaka for his experimental data, $\beta = 1.64 \pm 0.29$ and $M = 1.20 \pm 0.06$, the corresponding values of λ_c in our high-velocity experiments are of the order of 10 to 100 m. This is close to the amount of total displacement, but orders of magnitude higher than the grain size, the sample size or the sample surface roughness, which highlights the fact that in our high-velocity experiments, there is no obvious correlations between G_c (or D_c) and any characteristic length scale of the sample or the surface roughness. This shows that when slipping velocity and displacement are high enough, the initial surface roughness becomes unimportant with regard to the global mechanical behavior.

[26] Figure 5b shows that D_c decreases with increasing normal stress σ_n . However, the range of normal stresses investigated in our experiments is very narrow and is not comparable to normal stresses applied on real fault during natural earthquakes, even at shallow depths. Here we present scaling arguments that may explain the normal stress dependency of D_c and may allow extrapolating this parameter to higher normal stress. Considering that the mechanical work is all converted into heat and neglecting heat transport, the energy balance per unit of fault area leads to [*Rice*, 1999; *Di Toro et al.*, 2005; *Rice*, 2006; *Nielsen et al.*, 2008]

$$\bar{\tau}\Delta d = w\rho c\Delta T \tag{7}$$



Figure 11. TEM micrograph and electron diffraction patterns from material within the slipping zone. Numbers indicated on the TEM image each correlate with a diffraction pattern. The diaphragm used for diffraction was ~ 100 nm in diameter (i.e., corresponding to an area of 0.032 μ m²). The first diffraction pattern displays a diffusion ring pattern which reveals that the material is amorphous in this zone. A diffusion ring is still present on patterns 2 and 3 but is this time associated with diffraction spots, which indicates the presence of nanograins within an amorphous matrix. Patterns 4 and 5 show respectively the edge and the inner part of a well-crystallized metal oxide grain. Bragg fringes can be distinguished within the grain on the TEM image itself.

where $\bar{\tau}$ is the mean shear stress on fault, Δd is an increment of slip, w is the thickness of the slipping zone, ρc is the volumetric heat capacity of the rock and ΔT an increment of temperature. Now we assume that w is the affected thermal width, $w = \sqrt{\pi \alpha \Delta t}$, where α is the thermal diffusivity and Δt is the duration of slip. This assumption basically means that the thickness of the deformation zone increases as temperature diffuses. Now recalling that the slip velocity is $V = \Delta d/\Delta t$ and that the shear stress is proportional to the applied normal stress (Figure 5a), i.e., $\bar{\tau} = \bar{\mu} \sigma_n$, where $\bar{\mu}$ is the average friction coefficient, we get

$$\Delta d = \frac{\pi \alpha}{V} \left(\frac{\rho c \Delta T}{\bar{\mu} \sigma_n} \right)^2 \tag{8}$$

Equation (8) relies on the assumption of weakening due to shear heating. For such processes, equation (8) implies that dimensionally, the distance needed to decrease the shear

resistance of the fault should vary inversely with slip rate Vand with the square of the normal stress σ_n^2 . Taking representative values for the parameters in our experiments, such as $\Delta T = 400^{\circ}$ C, $\rho c = 1$ MPa °C⁻¹, $\bar{\mu} = 0.4$, $\sigma_n = 1$ MPa, V = 1 m s⁻¹ and $\alpha = 1$ mm² s⁻¹, we get an order of magnitude of several meters for Δd , which is very consistent with the data obtained in the HVF experiments. Figure 5b displays a fit using such dependence of D_c on σ_n . Now taking into account the derived dependency of D_c on normal stress, a value of 4 m at around 1 MPa normal stress should turn into about 4 cm to 0.4 mm at 10 to 100 MPa normal stress respectively, if the physical processes are still the same at those pressures. These values are much lower than the seismologically inferred slip weakening distances. However, the values measured in our experiments are for dry friction, on small samples (a few centimeters thickness), with a very thin gouge layer (around one millimeter). Largescale mechanical effects, such as off-fault dynamic yielding [Andrews, 1976, 2005], fault surface roughness [Scholz,

Experiment	μ_{p}	σ_n (MPa)	$G_c (MJ m^{-2})$	$H (MJ m^{-2})$	$E_{\text{total}} (\text{MJ m}^{-2})$
HVR780	0.85	1.31	5.0	7.9	12.9
HVR781	0.77	0.99	5.8	6.4	12.2
HVR782	1.02	0.61	5.5	6.3	11.8
HVR786	0.89	0.89	3.9	8.2	12.1
HVR787	1.09	0.51	3.8	6.7	10.5
HVR788	0.82	1.19	5.6	5.6	11.2

 Table 2. Energy Input During Experiments^a

 ${}^{a}G_{c}$ is the area under the slip-weakening curve, between the peak friction and the residual friction; *H* is the work of the residual friction; *E*_{total} is the total mechanical energy input, i.e., G + H.

1988] or gouge thickness [*Marone and Kilgore*, 1993] may play an important role in determining the characteristic slip weakening distance in real earthquakes. Moreover, the effects of pore pressurization due to shear heating have already been shown to influence the value of D_c in the MTL fault gouge [*Wibberley and Shimamoto*, 2005].

4.2. Energy Budget

[27] Since the values of shear stress and total displacement on the fault surface are measured for each run, the total mechanical energy input in each experiment can be directly estimated. Table 2 displays the values obtained in our experiments. The order of magnitude of total mechanical energy is about 10^7 J m⁻². Following the classical seismological view, two types of energy can de distinguished (see Figure 4): the fracture energy *G_c* and the heating energy *H*. We assume that all this energy is dissipated within the sample assembly, which implies that the energy losses in other parts of the apparatus or in the surrounding air are neglected. In this section, we estimate the amount of energy involved in (1) grain size reduction (a possible proxy for fracture energy), (2) heat production, and (3) dehydration and amorphization. Note that we will not tackle the problem

of radiated energy, as it is presently impossible to determine precisely the amount of energy radiated acoustically during HVF tests.

4.2.1. Estimate of the Surface Energy

[28] Let us consider a set of spherical grains of homogeneous size (initial diameter A_i). Within a thickness w (per unit of fault surface) the grains are fractured and their size is homogeneously reduced to a final diameter A_f . This simplified hypothesis leads directly to an estimation of an upper bound of the energy involved in grain size reduction; indeed, our microstructural analysis highlighted that all the grains have not been fractured down to same diameter. The number of grains per unit volume can be estimated by $N_{i,f} = (6/\pi)A_{i,f}^{-3}$, assuming that there is no void in the layer. The total surface area of the grains is given by $S_{i,f} = \pi w N_{i,f} A_{i,f}^2 S_{fault}$. Thus, the newly created surface area $\Delta S = S_f - S_i$ can be simply expressed as follows:

$$\frac{\Delta S}{S_{\text{fault}}} = 6w \left(\frac{1}{A_{\text{f}}} - \frac{1}{A_{\text{i}}} \right) \tag{9}$$

where 6 is a geometrical factor corresponding to the spherical geometry of the grains assumed here. In general, grain size reduction is such that $A_i \gg A_f$. Thus,

$$E_{\text{fracture}} \approx \gamma \times \frac{6w}{A_{\text{f}}} \qquad (\text{J m}^{-2})$$
 (10)

where γ is the average specific surface energy of the minerals. We used an average value of $\gamma \sim 1 \text{ J m}^{-2}$ (quartz at 500°C, from *Darot et al.* [1985]).

[29] Figure 12 displays the energy dissipated in grain size reduction as a function of the slipping zone thickness w, for two different final grain sizes. Considering a slipping zone thickness of about 10 μ m and a final diameter of the grains



Figure 12. Energy absorbed in grain size reduction, amorphization, and dehydration. Solid lines give the energy needed to dehydrate kaolinite (1 or 10 wt %). Dashed lines give the energy needed to reduce grain size, according to equation (10). Numbers in parentheses are the final grain diameters. Dotted lines give two estimates of the amorphization energy for quartz and feldspar: a lower bound (minus) and an upper bound (plus). An average value of fracture energy G_c calculated from the mechanical curves is given as a comparison. The gray area shows the range of slipping zone thicknesses observed in thin sections after the HVF experiments.



Figure 13. Numerically calculated temperature in samples. All the mechanical energy is assumed to be converted into heat on the slipping surface. Two experiments were simulated, HVR754 and HVR780. Values for normal stress and total displacement for each experiment are reported in the boxes. Parameter values are as follows: w = 0.5 mm, $\alpha_{\text{th}} = 1 \text{ mm}^2 \text{ s}^{-1}$, $\rho c = 2.7 \text{ MPa}^\circ \text{C}^{-1}$ and $v_{eq} = 1.03 \text{ m s}^{-1}$. (a and c) Spatial distribution of temperature after 40 s of slip. The temperature is higher on the edges, where the slip velocity is maximum. (b and d) Time evolution of temperature on the slipping surface at various radial positions. Maximum temperature reaches about +320°C for experiment HVR754 and about +470°C for experiment HVR780. These results are consistent with experimental data obtained by *Mizoguchi* [2005]. (e) Sketch showing the location and orientation of the finite volumes grid.

of only 1 nm, the energy absorbed into surface creation is around 10^5 J m⁻², i.e., only 2 to 5% of the fracture energy calculated from integration of the area under the mechanical curves. Values of fracture energy in Table 2 imply slipping zones of thickness \sim 200 to 500 μ m with average grain size of ~ 1 nm, which is clearly not observed. From this observation, we conclude that the mechanical energy dissipated in grain size reduction is negligible compared to the fracture energy G_c . This result is consistent with two recent studies performed by Ma et al. [2006] on samples from Chelungpu fault drilling and by Pittarello et al. [2008] on an exhumed pseudotachylyte-bearing fault zone in Italy. Ma et al. [2006] present a detailed microstructural analysis showing that the surface energy involved in grain size reduction within the slipping zone of the Chi-chi earthquake only represents $\sim 6\%$ of the fracture energy observed by seismologists, and Pittarello et al. [2008] also show that the

energy involved in grain size reduction and microcracking is only 3% of the total mechanical work during an earthquake. However, for natural faults, Figure 12 highlights the fact that the surface energy absorbed in grain size reduction is a limiting factor for the thickness of the slipping zone, which can only be from a few millimeters to centimeters in magnitude. However, note that an additional and likely nonnegligible part of the fracture energy might be absorbed in creating microcracks within the damage and process zones aside and ahead, respectively, of the fault [*Chester et al.*, 2005].

4.2.2. Heat Production

[30] To quantify the increase in temperature within the specimens due to heat production, we integrated numerically the energy converted into heat. A simple two-dimensional finite difference diffusion model was used. Figure 13e displays the orientation of the finite difference grid. The

sample was considered to be homogeneous since thermal properties of gouge and gabbro are similar. The boundary conditions were assumed to be adiabatic, because of the presence of the Teflon ring. This assumption leads to an upper bound for the final temperature since in the experiments there are heat losses due to conduction and radiation of heat around the sample. The evolution of shear stress with time was approximated by the empirical equation (1):

$$\tau(t) = \tau_0 \exp\left(\frac{\ln(0.05)v_{\rm eq}t}{D_c}\right) + \tau_{\rm r}$$

For the sake of simplicity, we assume here that all the mechanical energy is converted into heat. This assumption seems relevant in the light of the discussion above, although it may induce a slight overestimate of the temperature because of heat sinks such as dehydration and amorphization. The heat flux is given by

$$Q(r,t) = 0.5\tau\omega r \tag{11}$$

where ω is the angular velocity, and *r* the radius from the center of the axis of rotation. The heat equation can thus be written as follows:

$$\frac{\partial T}{\partial t} - \frac{Q(r,t)}{w\rho c} = \alpha_{\rm th} \left(\frac{\partial^2 T}{\partial y^2} + \frac{\partial^2 T}{\partial r^2} \right) \tag{12}$$

with *w* being the width of the heat production zone, ρ the density, *c* the specific heat capacity and $\alpha_{\rm th}$ thermal diffusivity. Our model is close to a model with a heat flux boundary condition on the gouge surface [*Carslaw and Jaeger*, 1959]. Equation (12) was numerically integrated using the following parameter values: w = 0.5 mm, $\alpha_{\rm th} = 1$ mm² s⁻¹, $\rho c = 2.7$ MPa °C⁻¹ and $v_{eq} = 1.03$ m s⁻¹. A thickness of 0.5 mm was chosen because it corresponds roughly to half of the real gouge thickness in our experiments. Several simulations with varying thickness around this value, from 0.1 to 1 mm, show that the temperature increase is not significantly influenced by this parameter with such a small value.

[31] The increase in temperature after 40 s of slip depends on the experimental conditions. On the fault plane, it reaches about +320°C when the normal stress is 0.60 MPa, and about +470°C when normal stress is 1.32 MPa. These values are consistent with the data previously obtained in similar experiments, using thermocouples in holes inside the stationary part of the host rock [Mizoguchi, 2005], and are also consistent with our microstructural observations that reveal no evidence of melting within the sheared gouge. However, this relatively small increase in temperature (less than 500°C) seems inconsistent with kaolinite dehydration, which usually occurs a higher temperatures, starting at around 500°C [e.g., Frost et al., 2004; L'vov and Ugolkov, 2005]. Yet our data clearly indicates amorphization and humidity increase during shearing. Two nonexclusive hypotheses can be formed: (1) kaolinite is mechanically amorphized and only surface adsorbed water is released during the experiment; and (2) kaolinite is really dehydrated, i.e., thermally decomposed. The latter scenario can be considered as realistic since it has been shown [Makó et al., 2001; Horváth et al., 2003] that kaolinite milled with harder grains such as quartz, dehydrates at lower temperatures than anticipated. In the same way, the dramatic grain size reduction within the slipping zone implies an increase of kaolinite surface area, and may thus influence the dehydration reaction kinetics. Furthermore, flash heating may also occur at asperity contacts between grains within the slipping zone, and local flash temperatures at contacts can be much higher than the average fault temperature [Archard, 1958/1959; Rice, 2006]. Similar explanations are also given by Hirose and Bystricky [2007], who observed water vapor release from serpentinite sheared at high slip rates, but measured temperatures lower than the equilibrium temperature of dehydration. We emphasize that although our numerical integration draws no conclusive and definitive proof that kaolinite is dehydrating at high slip rates in our experiments, our mineralogical data and relative humidity data seem to point in this direction.

4.2.3. Amorphization and Dehydration

[32] Another energy sink in our experiment is that of mineral amorphization within the gouge. We estimate the amorphization energy of some minerals contained in the gouge: the enthalpy change at 25°C from crystalline quartz and feldspar to glass ranges from the order of 5 kJ mol⁻¹ for quartz (60 g mol⁻¹ molecular weight) to 50 kJ mol⁻¹ for feldspar (278 g mol⁻¹ molecular weight) [*Robie et al.*, 1979], i.e., around 80 J g⁻¹ for quartz to 200 J g⁻¹ for feldspar. With a thickness *w*, and an average density of 2700 kg m⁻³, the energy absorbed by mineral amorphization $E_{\rm am}$ can be bounded as follows:

$$2.2 \times 10^8 w \le E_{\rm am} \le 5.4 \times 10^8 w$$
 J m⁻² (13)

In the same way, we estimate the energy needed to dehydrate kaolinite E_{dhy} per unit of fault area: the enthalpy change of dehydration is of the order of 1000 kJ mol⁻¹ [*L'vov and Ugolkov*, 2005]. The molecular weight of kaolinite is 258 g mol⁻¹, its density is about 2700 kg m⁻³. For a mass fraction of kaolinite f_{kaol} and a thickness *w*, we get

$$\begin{split} E_{\rm dhy} &\approx w \times f_{\rm kaol} \times 2700/0.258 \times 10^6 {\rm J} {\rm m}^{-2} \\ &\approx w \times f_{\rm kaol} \times 1.1 \times 10^{10} {\rm J} {\rm m}^{-2} \end{split}$$

[33] These values are plotted against thickness w in Figure 12 (solid and dotted lines). The measured thickness of the slipping zone ranges from 2 to 40 μ m. For both mineral amorphization and kaolinite dehydration, the corresponding energy values are much lower than the calculated fracture energy G_c (and thus lower than the total mechanical energy input). Importantly, this clearly indicates that kaolinite dehydration is a negligible energy sink compared to the total energy input. Let us define E_{heat} the energy converted into heat within the slipping zone per unit of fault area. This energy can be expressed as

$$E_{\text{heat}} = c_p \rho w (T_{\text{final}} - T_{\text{initial}})$$

where c_p is the specific heat capacity (around 1000 J (kg K)⁻¹ in most rocks), $(T_{\text{final}} - T_{\text{initial}})$ the temperature increase (around 400°C following Figure 13) and ρ is the density of

 Table 3. Mineral and Porosity Data of Fault Gouges^a

Fault Zone	Hydrous Minerals (wt %)	Total Water Content (wt %)	Porosity
MTL SAFOD	Kaolinite (~10-20%)	1.2-2.4%	0.04 ^b
-1360 m	illite, chlorite, I-S (4-15%), laumontite (6-13%)	1.5-4.2%	0.05 ^c
—1926 m	illite, chlorite $(1-6\%)$, laumontite $(1-4\%)$	0.3 - 1.5%	0.05 ^c
-2545 m	chlorite, illite (13-30%), I-S (9-14%)	3.3-6.6%	0.05 ^c
-3067 m	illite (48-51%), I-S (14-18%)	9.3-10.4%	0.05°
-3300 m	chlorite, illite, I-S (14-60%)	2.1-9%	0.05°
Aegion	chlorite (2-3%), illite (8-12%)	0.9-1.4%	0.08^{d}

^aMineral data of MTL fault gouge were estimated from X-ray diffraction patterns previously detailed (section 2.1); San Andreas fault data from *Solum et al.* [2006]; Aegion fault data from *Sulem et al.* [2004]. Total water content is an estimate of the sum of the water amount in each hydrous mineral: illite \approx 12%, chlorite \approx 15%, I-S (illite-smectite) \approx 15%, laumontite \approx 15% and Kaolinite \approx 14%.

^bPorosity data from *Wibberley and Shimamoto* [2003].

^cPorosity for San Andreas fault rocks was assumed to be 5%, because of the lack of data, a value chosen such that its order of magnitude is consistent with other measured data.

^dSulem et al. [2007] (porosity at high confining pressure).

the rock (about 2700 kg m⁻³ for most rocks). This yields $E_{\text{heat}} \approx w \times 1.1 \times 10^9$ J m⁻². Thus, the ratio $E_{\text{dhy}}/E_{\text{heat}}$ is of the order of $10 \times f_{\text{kaol}}$, leading to a value of about 1 in the case of 10% mass fraction of kaolinite (Table 3). This shows that the dehydration energy is comparable to the heating energy in the slipping zone itself. However, the width of the heated part of the sample is much larger than the width of the slipping zone (Figure 13), which implies that the dehydration energy is by far less than the total heating energy in the whole sample. These estimates are very rough but seem to demonstrate that in our experiments, the major part of G_c is converted either into heat or radiated energy.

[34] Note that Figure 12 also indicates that for a natural fault, the width of the slipping zone must be bounded both by the energy input and the physical processes occurring within it (grain size reduction, amorphization, dehydration, melting, etc.). For a given total energy input of an earth-quake (depending on its magnitude), the maximum thickness of the ultralocalized deformation zone depends on the degree of grain size reduction, the extent of amorphization or partial melting and/or other thermal or chemical couplings (such as dehydration). This kind of argument could be useful to give additional constraints on the width of the slipping zone in natural fault zones, although the amount of comminution or grain crushing may be less for broader slip zones.

[35] Despite these considerations, the processes that induce slip weakening during HVF experiments are still not clear. However, our results suggest that the main thermodynamic variable that influences weakening during our experiments is likely to be temperature. Our analysis has shown that most of the mechanical energy seems to be converted into heat (and radiated energy?) and that a change in the mechanical experimental conditions (normal stress or slip velocity) directly affects both heat production (i.e., the temperature on the slipping plane) and the value of D_c . Thus the weakening mechanism in our experiments may be thermally activated. This is consistent with the extrapolation of D_c values described in section 4.1. In our experiments, thermal pressurization of pore fluid is probably a negligible weakening phenomenon because hydraulic diffusivity is high (the gouge is not consolidated and the pore fluid is in gaseous form), which is certainly not the case in nature

[e.g., Sibson, 1973; Lachenbruch, 1980; Wibberley and Shimamoto, 2005; Noda and Shimamoto, 2005; Rice, 2006].

5. Consequences of Coseismic Dehydration on Faults

[36] Thermal pressurization of pore fluids is an important thermohydromechanical coupling that takes place on a fault during seismic slip, which involves a pore pressure buildup induced by frictional heating within a fault zone. It can be modelled by coupling heat diffusion equations and poroelasticity equations and exhaustive theoretical works have already been published [e.g., *Sibson*, 1973; *Lachenbruch*, 1980; *Mase and Smith*, 1985; *Andrews*, 2002; *Rice*, 2006]. Likewise, dehydration of hydrous minerals may also trigger pore pressure increase. We estimate the pore pressure increase due to thermal dehydration of hydrous minerals (such as clays), and compare it to theoretical predictions of the pore pressure increase due to thermal pressurization of pore fluids.

[37] Let us consider a fault rock containing hydrous minerals, assumed to be a porous, fluid-saturated medium. Using an adiabatic, undrained approximation [e.g., *Lachenbruch*, 1980; *Rice*, 2006], the increase in fluid pressure due to thermal pressurization can be estimated by

$$\Delta p_{\rm thp} = \Lambda \Delta T_{\rm d} \tag{14}$$

where $\Delta T_{\rm d}$ denotes the temperature increase from the initial state when dehydration occurs, and Λ is the thermal pressurization coefficient. Λ is given by

$$\Lambda = \frac{\lambda_f - \lambda_n}{\beta_f + \beta_n} \tag{15}$$

where λ_f and β_f denote the thermal expansion coefficient and the isothermal compressibility of the pore fluid, respectively, and λ_n and β_n denote the thermal expansion coefficient and the isothermal compressibility of the pore space, respectively.

[38] The mass increment of water due to dehydration (source term dm_d) can be written

$$dm_{\rm d} = \rho \chi d\xi \tag{16}$$



Figure 14. \mathcal{D} values for several fault gouges. Solid and dashed lines correspond to different dehydration temperatures. Density ratio ρ/ρ_f is 2.8, and thermal expansion difference $(\lambda_f - \lambda_n)$ is $10^{-3\circ}$ K⁻¹. Data are reported in Table 3. The serpentinite case is plotted as an example, assuming 3% porosity and 15% water content. The horizontal length of the bars indicates an estimate of the errors in the ratio χ/n , and their vertical length indicates the range of temperatures used. Most of the corresponding \mathcal{D} values are at least close to 1, which shows that dehydration effects cannot be neglected.

where χ denotes the mass fraction of water that can be released during dehydration, ρ is the average rock density and ξ is the extent of reaction. This linear relationship between dm_d and $d\xi$ relies on the hypothesis that water is constantly released during the dehydration process. Note here that equation (16) depends neither on the dehydration mechanism nor on the type of fluid that is expelled from the mineral structure: the only assumption is that fluid is being released from the minerals. In particular, the fluid can be adsorbed water as well as bonded water. In the following the term "dehydration reaction" will refer to either of these cases.

[39] The dehydration term (right-hand side of equation (16)) depends on (1) the amount of water that can be released χ and (2) the reaction kinetics $(\partial \xi / \partial t)$. Previous studies on thermal pressurization have shown that temperature can increase very rapidly close to the fault plane (or within the deformation zone) [e.g., *Rice*, 2006; *Rempel and Rice*, 2006]. Such short timescales imply that temperature can well overshoot the thermodynamic equilibrium temperature of the dehydration reaction, which will increase the driving force for dehydration. Thus, the reaction is considered here as instantaneous at a threshold temperature denoted $T_{\rm d}$. In consequence, the effect of dehydration will be a pore pressure peak slowly decreasing by diffusion of the pore fluid.

[40] For the sake of simplicity, we assume no significant change of porous volume due to the dehydration reaction.

The additional pore pressure source term can then be expressed as follows:

$$\Delta p_{\rm d} = \chi \frac{\rho}{\rho_f} \frac{1}{\beta^*} \tag{17}$$

where ρ_f denotes the density of water and β^* the storage capacity of the porous medium. β^* is defined by $\beta^* = n(\beta_n + \beta_n)$ β_{f} , with *n* being the rock porosity, β_{n} the isothermal pressure expansivity of the pore space and β_f the isothermal compressibility of the pore fluid. For example and using a set of representative values for the various parameters: $\chi = 1\%$ of total mass fraction is water, $\rho/\rho_f =$ 2.8, $\beta^* = 10^{-10}$ Pa⁻¹, the pore pressure rise due to the reaction is $\Delta p_{\rm d} = 280$ MPa, which is of the order of the overburden pressure in the first ten kilometers of the crust. This has an important consequence, being that if the initial pore pressure is high enough, the sum of overpressures due to (1) thermal pressurization and (2) dehydration reactions can overshoot the normal stress applied to the fault. Such overpressures do not occur in the case of thermal pressurization only, which leads to an asymptotic increase of pore pressure up to the normal stress applied on the fault. In the case of overpressure due to dehydration, the fault would be subjected to tensile stress. Because rocks have low tensile strength, it is likely that hydraulic fracturing would take place within the fault walls and fault process zone.

[41] From the expression for peak pressure at the time of dehydration Δp_d (equation (17)), a nondimensional number denoted \mathcal{D} can be defined by

$$\mathcal{D} = \frac{\Delta p_{\rm d}}{\Delta p_{\rm thp}} = \frac{\chi}{n} \frac{\rho}{\rho_f} \frac{1}{(\lambda_f - \lambda_n) \Delta T_{\rm d}}$$
(18)

 \mathcal{D} compares the relative importance of the pore pressure terms due to dehydration and thermal pressurization. When $\mathcal{D} \ll 1$, dehydration becomes a negligible phenomenon compared to thermal pressurization; if $\mathcal{D} \sim 1$ then both effects are equally important. If $\mathcal{D} \gg 1$ then dehydration is dominating and hydraulic fracturing might be observed. However, let us emphasize once again that the dehydration reaction can only occur after a sufficient temperature increase (up to the threshold T_d), and thus, thermal pressurization always prevails at first, up to the temperature threshold. Figure 14 plots the values obtained for \mathcal{D} as a function of the ratio χ/n . This ratio compares the relative importance of the amount of water released with the available pore space. Values of this ratio can be estimated for several fault zones, and the parameter values used in the estimation are presented in Table 3. One fault gouge sample recovered from SAFOD exhibit a low \mathcal{D} value (close to 0.1), but most measurements on other SAFOD fault zones give higher values: \mathcal{D} ranges from 1 (e.g., Aegion or MTL fault gouges) to 10 (SAFOD). Dehydration effects are thus likely to be nonnegligible in most of the fault zones reported here. This is due to the nearly ubiquitous presence of hydrous minerals, mainly clays, in superficial and medium depth faults. However, as reported by Solum et al. [2006], the amount of clay is highly variable from one fault gouge to another.

[42] Likewise, serpentinized rocks usually have very low porosity and their hydrous minerals content is very high, and it appears that dehydration is a phenomenon that has to be taken into account when considering earthquake propagation within these materials. Moreover, in the case when $\mathcal{D} \gg 1$, the pore pressure increase due to dehydration may dramatically reduce the effective stress, and thus prevent further frictional heating. Therefore dehydration reactions limit the increase in temperature within the slipping zone, not so much because they are generally endothermic, but because of the reduction in effective normal stress. This may be an additional explanation for the apparent scarcity of pseudotachylytes in nature: if the amount of hydrous minerals present in a fault zone is high enough, melting cannot occur easily since dehydration occurs first and stops (or slows down) heat production. We find here agreement with the assertion that pseudotachylytes can only be produced in dry, crystalline rocks, as reported by Sibson and Toy [2006].

6. Conclusions

[43] The main conclusions of our study are as follows:

[44] 1. Field data on the MTL (Tsukide outcrop) show several gouge layers with varying kaolinite content which might be interpreted as a result of kaolinite decomposition during a seismic event.

[45] 2. HVF tests on MTL gouge samples show that shear stress decreases abruptly during slip. The characteristic slip distance D_c ranges from a few meters to a few tens of meters

and decreases with normal stress. The D_c values do not scale with any geometric length scale of the samples. To compare these values to seismologically inferred values, an extrapolation of D_c to higher normal stresses can be performed assuming that D_c is controlled by heating within the gouge sample. This shows that D_c varies with the inverse of slip velocity and the inverse of the square of the normal stress, leading to very small values from around 4 cm to 0.4 mm at 10 to 100 MPa normal stress, respectively. However, such analysis relies on the assumption that the thickness of the slipping zone is the thermally affected width and does not take into account the influence of the presence of fluids and large-scale geometric and mechanical effects. These may play an important role in the effective value of D_c parameter in real earthquakes.

[46] 3. SEM and TEM investigations of our deformed gouge highlight the presence of a thin, highly deformed zone, interpreted as the main slipping zone. The siliceous and aluminous fraction of this zone are amorphized, and contains nanometric grains. Only a few metal oxides are preserved. The microstructure is deeply altered during deformation with such intensity that it must be related to the temperature increase and to the mechanical interaction of grains.

[47] 4. Investigations of the energy balance show that the surface energy created during shearing corresponds to only 2 to 5% of the fracture energy calculated from the mechanical data. From this observation, we conclude that the mechanical energy dissipated in comminution is negligible compared to the fracture energy G_c . Moreover, we showed that kaolinite dehydration and mineral amorphization are also negligible energy sinks compared to the total energy input. These estimates are very rough, but seem to indicate that in our experiments at least, the major part of G_c was converted either into heat or into radiated energy. However, for a given total energy input (depending on the magnitude of the earthquake), the maximum thickness of the ultralocalized deformation zone depends on the amount of grain size reduction, the extent of amorphization or partial melting and/or other thermal or chemical couplings (such as dehydration). This kind of argument could be useful to give additional constraints on the width of the slipping zone in natural fault zones.

[48] 5. Our experimental work shows that kaolinite may dehydrate seismic slip velocities. This has important largescale mechanical implications. Indeed, coseismic dehydration of clays (or hydrous minerals in general) may trigger a pore pressure pulse, superimposed with the pressure increase from thermal pressurization of pore fluid. For water contents as low as 1% (which is the case of some of the SAFOD cores, for instance), this pulse can be of the order of the overburden pressure of the first ten kilometers of the crust. In consequence, if the initial pore pressure is high enough, the sum of overpressures due to (1) thermal pressurization and (2) dehydration reactions can exceed the normal stress applied to the fault. In such a case, frictional heating can slow down or even stop, except if the fault walls are strongly damaged, which may induce a rapid drainage of the gouge and increase the frictional heating. In any case, this effect may strongly affect the thermal evolution of the fault, and further investigation are needed to constrain which phenomenon is more likely to occur (increase or

decrease in frictional heating). If it turns out that damage is not important, then coseismic dehydrations might be a plausible explanation for the apparent scarcity of pseudotachylytes in nature.

^[49] In the future, additional constraints should be added: (1) by performing systematic mineralogical analysis on natural fault gouges, (2) by constructing numerical models that incorporate an additional fluid source term in the thermal pressurization equations coming from dehydration, and (3) by investigating experimentally fluid pressurization during dehydration of hydrous minerals, especially the corresponding irreversible deformation.

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