Variation of Hydraulic Properties Due to Dynamic Fracture Damage: Implications for Fault Zones

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Abstract

High strain rate loading causes pervasive dynamic microfracturing in crystalline materials, with dynamic pulverization being the extreme end-member. Hydraulic properties (permeability, porosity, and storage capacity) are primarily controlled by fracture damage and will therefore change significantly by intense dynamic fracturing—by how much is currently unknown. Dynamic fracture damage observed in the damage zones of seismic faults is thought to originate from dynamic stresses near the earthquake rupture tip. This implies that during an earthquake, hydraulic properties in the damage zone change early. The immediate effect this has on fluid-driven coseismic slip processes following the rupture, and on postseismic and interseismic fault zone processes, is not yet clear. Here, we present hydraulic properties measured on the full range of dynamic fracture damage up to dynamic pulverization. Dynamic damage was induced in quartz-monzonite samples by performing uniaxial high strain rate (> 10^6 s^{-1}) experiments in compression using a split-Hopkinson pressure bar. Hydraulic properties were measured on samples subjected to single and successive loadings, the latter to simulate cumulative damage from repeated rupture events. We show that permeability increases by 6 orders of magnitude and porosity by 15% with dissipated energy up to dynamic pulverization, for both single and successive loadings. We present damage zone permeability profiles induced by earthquake rupture and how it evolves with repeated ruptures. We propose that the enhanced hydraulic properties measured for pulverized rock decrease the efficiency of thermal pressurization, when emplaced adjacent to the principal slip zone.

1. Introduction

Brittle deformation of low-porosity crystalline rock in compression is accommodated by the nucleation and growth of microfractures from preexisting flaws (e.g., pores and grain boundaries) and their coalescence into larger fractures along which the material fails (Paterson & Wong, 2005). At low (quasi-static) strain rates, these steps toward failure occur more or less sequentially. At high (dynamic) strain rates, the rate at which stress on the rock increases approaches the rate at which each step occurs (e.g., fracture growth and microfracture coalescence) so that stress increases at a faster rate than microfractures can form. This results in more stored elastic strain energy, an increase in apparent fracture toughness, and an increase in macroscopic strength (Bhat et al., 2012; Kolsky, 1949; Liu et al., 1998; Zhang & Zhao, 2013a). Hence, the steps to failure occur in parallel, that is, a larger amount of microfractures grow and coalesce concurrently with the propagation of coalesced larger fractures (Hild et al., 2003). The transition from quasi-static fracturing to dynamic fracturing typically occurs at strain rates of 10^6–10^5 s^{-1} for low-porosity crystalline rock (Zhang & Zhao, 2013b). At even higher strain rate, dynamic fracturing leads to a sharp decrease in postloading fragment size (i.e., the rock pulverizes), according to theoretical models (Grady, 1982; Glenn & Chudnovsky, 1986; Hild et al., 2003). Experiments have demonstrated such pulverization of crystalline rock above a strain rate threshold of 150–200 s^{-1} under uniaxial compression (Doan & Gary, 2009; Xia et al., 2008; Yuan et al., 2011). Pulverization textures are encountered in many geological settings, suggesting that brittle high strain rate deformation occurs in crustal-scale strike-slip fault zones (Dor et al., 2006; Doan & Gary, 2009), near impact cratering events (e.g., Grady & Kipp, 1987; Melosh, 1984), and during explosive volcanic eruptions where both magma and volcanic host rock are fragmented and expelled as volcanic ash (e.g., Haug et al., 2011). Pulverization textures are encountered in many geological settings, suggesting that brittle high strain rate deformation occurs in crustal-scale strike-slip fault zones (Dor et al., 2006; Doan & Gary, 2009), near impact cratering events (e.g., Grady & Kipp, 1987; Melosh, 1984), and during explosive volcanic eruptions where both magma and volcanic host rock are fragmented and expelled as volcanic ash (e.g., Haug et al., 2011). 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Figure 1. Schematic fault damage zone structure with the peak strain rate caused by dynamic rupture along the fault core. The peak strain rate decays with distance from the fault. Damage zone rock close to the fault experiences strain rates above the pulverization threshold ($10^2$ s$^{-1}$ in compression (Aben et al., 2016; Doan & Gary, 2009), $10^1$ s$^{-1}$ for isotropic tension (Griffith et al., 2018)); further from the fault strain rates are sufficient for dynamic fracturing. Quasi-static failure may occur at even larger distance from the fault, or dynamic stresses fall below the quasi-static peak strength of the rock. Figure modified from Aben et al. (2016).

In fault zones, off-fault damage consists of a combination of high strain rate and quasi-static damage accrued and overprinted during the fault’s history. Quasi-static loading mechanisms constitute fault growth, fault wear, and interaction between faults (Mitchell & Faulkner, 2009), whereas dynamic fracturing is imposed by the stress field of a propagating earthquake rupture (Freund, 1979; Poliakov et al., 2002; Reches & Dewers, 2005) (Figure 1). Separating the coseismic dynamic fracture contribution from the quasi-static one is challenging, except when pulverized damage zone rock is present. The key characteristic of pulverized damage zone rock is their lack of shear or rotation of fragments—leading to preservation of its original texture (Dor et al., 2006; Mitchell et al., 2011), while they are fragmented down to the micron scale (Dor et al., 2006; Muto et al., 2015) and they disintegrate upon light touch (Dor et al., 2006; Mitchell et al., 2011). High strain rate loading experiments in compression recreated pulverization textures above a strain rate threshold of about $200$ s$^{-1}$, whereby the lack of shear is explained by a short loading duration (Doan & Gary, 2009). The strain rate threshold increases as a function of confining pressure (Yuan et al., 2011) and decreases by progressive accumulation of fracture damage through successive high strain rate loadings (Aben et al., 2016; Doan & D’Hour, 2012). Other loading mechanisms proposed for off-fault pulverization, such as tensile dynamic loading (Griffith et al., 2018) and decompression of pore fluid saturated rock (Mitchell et al., 2013), rely on strain rates that are about 2 orders of magnitude lower ($1$ s$^{-1}$) than those necessary for pulverization in compression but still far exceed conventional laboratory deformation strain rates ($10^{-5}$ s$^{-1}$) and tectonic strain rates. Pulverization in compression is expected to occur in a narrow band (at most a few meters wide) around the fault core, based on strain rate predictions from rupture models (Aben et al., 2017a; Griffith et al., 2018; Xu & Ben-zion, 2017) (Figure 1). It is important to note that high strain rate fracture damage is expected outside the pulverized band as well (Aben et al., 2016) (Figure 1), although difficult to distinguish from quasi-static damage when complete overprinting has occurred. Nonetheless, the fracture damage in fault damage zones will have a dynamic character.

Fluid flow properties in low-porosity crystalline rock (e.g., permeability, porosity, and storage capacity) are primarily controlled by fracture damage (Bernabé et al., 2003). It has been shown that fracture distributions resulting from dynamic and quasi-static deformation vary distinctly (Fondriest et al., 2017; Muto et al., 2015); despite this, little is known about the hydraulic properties of dynamically fractured rock. The evolution of fluid flow properties during and after quasi-static deformation has been studied extensively.
Hydraulic properties have been measured on sheared fault core materials (e.g., Boutreau et al., 2008; Mizoguchi et al., 2008; Wibberley, 2002) and damage zone rock (e.g., Allen et al., 2017; Mizoguchi et al., 2008; Rempe et al., 2018) sampled in natural fault zones; however, they are the product of a long history and contain quasi-statically and dynamically accrued damage. To date, no studies have systematically measured hydraulic properties of dynamically fractured and pulverized rock, thus obscuring the hydraulic structure of fault damage zones that contain the entire spectrum of dynamic fracture damage (Figure 1). Yet dynamic fracture damage is pivotal in fluid-fault rock interaction during various stages of the seismic cycle, largely because it is located directly adjacent to the fault core. The fluid flow properties in the damage zone rock are changed by the stress field of the propagating earthquake rupture tip, which may occur at an early stage of the earthquake. Slip-weakening mechanisms driven by the presence of pore fluids, in particular thermal pressurization (Lachenbruch, 1980), initially depend on the fluid flow properties of the principal slip zone only. With larger slip, the permeability and hydraulic diffusivity of the damage zone wall rock begin to influence its efficiency (Rempel & Rice, 2006; Rice, 2006). In addition, near-instantaneous dilation near the rupture tip and in the fault damage zone causes a fluid pressure drop that may affect slip-weakening mechanisms (Brantut, 2020; Martin, 1980). At the postseismic and interseismic stages, the fault damage zone provides efficient pathways for the migration of hydrothermal fluids. These fluid have the potential to chemically alter fault cataclasites and reactive gouges in the fault core, thereby changing the fault strength and slip behavior (Kaduri et al., 2017; Lockner et al., 2011). Fractures participate in fluid flow for a limited time only as they close over a range of timescales by a variety of recovery and sealing mechanisms (Aben et al., 2017b; Brantley et al., 1990; Renard et al., 2000). Arguably, coseismic fracturing and wear are the only sources for reopening of sealed fractures or creation of new fractures and are thus pivotal in maintaining a high fault zone permeability relative to undamaged host rock. These examples illustrate the necessity for better constraints on the fluid flow properties in high strain rate-deformed low-porosity crystalline rock.

Here, we present the hydraulic properties across the entire range of dynamic fracture damage up to pulverized rocks. As dynamic fracture damage is accumulated over successive earthquake rupture events rather than a single event, we have performed single and successive high strain rate loading experiments. These experiments were performed in compression on quartz-monzonite, a low-porosity crystalline rock. We measured permeability, porosity, storage capacity, seismic velocities, and fracture densities on postmortem samples. We then discuss the results in relation to fault zones, with specific examples illustrating their impact on the evolution of fault damage zone permeability during and after an earthquake and on the efficiency of thermal pressurization.

2. Materials and Methods

A series of cylindrical samples were cored from a block of quartz-monzonite (for characterization of some petrophysical and mineralogical properties of this rock, see Aben et al., 2016), and the two end faces of each sample were ground parallel within 20–30 μm. The length/diameter ratio was close to 1 (15 mm by 15 mm) to eliminate inertia effects (Gama et al., 2004; Zhang & Zhao, 2013b). Radial dynamic confinement within the sample during loading was limited to at most 0.5 MPa for the range of applied strain rates (Aben et al., 2016; Forrestal et al., 2007). The samples were inserted into a tight rubber jacket to ensure postmortem recovery of the microstructure. The rubber jacket was left on during postmortem porosity and permeability characterization to minimize disturbance and slip along microfractures.

Uniaxial high strain rate loading conditions were applied by inserting the samples into a mini-split-Hopkinson pressure bar apparatus at ISTerre laboratory in Grenoble. The apparatus consists of steel input and output bars (2 cm diameter). The rock sample is placed between the two bars. A steel striker is launched toward the input bar by an air gun. A planar stress wave resulting from the striker impact travels the length of the input bar and loads the sample. Part of the input stress wave is reflected in opposite direction, and part is transmitted through the sample into the output bar. Strain gauges placed on the input and output bars monitor the propagation of the input, reflected, and transmitted waves. The full stress-strain history of the loading is obtained by applying a 1-D wave analysis on the strain gauge data after a dispersion correction for the 2-D radial symmetry of the bars. More details on the machine are described in Aben et al. (2016) and on the 1-D wave analysis in Aben et al. (2017a) and Gama et al. (2004).
Figure 2. (a) Incident stress waves measured as change in voltage by the strain gauges for the three sets with constant loading parameters. Striker length and airgun pressure ($P_{\text{air}}$) are given. (b) Stress-strain curves for three repeated loadings with a constant incident stress wave. Dissipated energy is defined as the area bounded by the stress-strain curve.

Twenty-six samples were split into two batches. One batch was subjected to single loadings with varying loading conditions. Samples from the other batch were subjected to successive loadings where each incident stress wave was similar to the previous loading. Three different loading conditions were used for the successive loading sample batch (Figure 2a). The use of pulse shapers dampened the initial stress rate, so that stress equilibrium along the length of the sample was achieved within the elastic loading interval. Nonetheless, error bars are used for the input and output stresses as well as for dissipated energy, which is computed from both the input and output stress and strain data. The loading conditions for all samples are summarized in Table 1. The critical strain rate was defined in two ways: (1) as the maximum strain rate before the brittle yield or failure point of the sample and (2) as the strain rate at the onset of yielding. This yield point in the loading history was defined as the onset of nonlinearity in the stress versus strain curve. High-speed camera footage confirms that this point represents the onset of fracture (Text S1 in the supporting information).

Porosity, permeability, specific storage capacity, and $P$ wave velocities were measured on the damaged samples and some intact samples. The microstructures of three samples were obtained using X-ray micro-computed tomography (X-ray $\mu$CT). Some samples from Aben et al. (2016) and Aben et al. (2017b) were included for analyses (see Table 1). $P$ wave velocity results are presented in Text S3 in the supporting information.

Porosity $\phi$ was obtained by measuring the volume of the solid only (the deformed sample and the jacket) in a helium gas pycnometer. The total volume, consisting of the solid rock and the pores, was calculated from calliper measurements (all chips fallen of the edges of the sample were included in the pore volume by assuming an initially solid cylinder). The volume fraction of connected pores is the difference between the total volume and the solid volume. Variability was about 0.5% porosity, because the jacket included in the measurement is not a perfect hollow cylinder. Although using the modified wax cloth method (Girty et al., 2008) would result in more accurate measurements, the samples would be unsuited for permeability measurements after due to the clogging of the pores. Since the jacket is included in all porosity measurements (for intact as well as fragile pulverized samples), the results are internally consistent.

Permeability $\kappa$ was measured under hydrostatic conditions in a servo-controlled steady-state flow permeameter ($10^{-12} > \kappa > 10^{-21}$ m$^2$) at UCL in London (Benson et al., 2005). The apparatus consists of a confining pressure vessel (up to 200 MPa) that houses the sample assembly. The sample was inserted into a rubber jacket, equipped with pore fluid distribution plates, and placed in the sample assembly between two end caps that allow pore fluids to flow in and out of the sample. The upstream and downstream pore fluid reservoirs are equipped with servo-controlled hydraulic intensifiers. The pressure in both reservoirs was set so that a constant pore fluid pressure difference was established along the sample, resulting in flow of the pore fluid (deionized water was used here). Once steady-state flow was achieved, the permeability was calculated from the pressure difference, the rate of fluid flow, and the sample dimensions, according to Darcy’s law.
Sample Loading Settings (Number of Loadings, Impact Striker Length, Compressed Air Gun Pressure), Postmortem Measurements ($\phi$ = Porosity, $\kappa$ = Permeability, $S$ = Storage Capacity, $V_P$ = P Wave Velocity (Text S3)) and Microstructural Analysis by X-ray $\mu$CT

| Sample # | Loading # | Striker length (cm) | $P_{\text{air}}$ (bar) | $\phi$ | $\kappa$ ($\Delta P_f$) (MPa) | $S$ | $V_P$ | X-ray $\mu$CT |
|----------|-----------|---------------------|------------------------|-------|-------------------------------|-----|------|-------------|
| QM1      | 1         | 15                  | 6.0                    | x     |                               |     |      |             |
| QM2      | 1         | 15                  | 4.0                    | x     |                               |     |      |             |
| QM3      | 1         | 15                  | 6.5                    | x     |                               |     |      |             |
| QM6      | 1         | 15                  | 5.6                    | x     |                               |     |      |             |
| QM8      | 1         | 15                  | 6.0                    | x     |                               |     |      |             |
| QM9      | 1         | 15                  | 5.5                    | x     |                               |     |      |             |
| QM10     | 1         | 15                  | 5.5                    | 0.2   | x                             | x   | x   |             |
| QM11     | 1         | 15                  | 5.0                    | x     | 0.5                           |     |      |             |
| QM12     | 1         | 15                  | 4.5                    | x     | 1.0                           |     | x   |             |
| QM13     | 1         | 15                  | 4.0                    | x     | 1.0                           |     | x   |             |
| QM14     | 3         | 15                  | 4.0                    | x     |                               |     |      |             |
| QM15     | 2         | 15                  | 4.0                    | x     |                               |     |      |             |
| QM16     | 3         | 15                  | 3.5                    | x     | 0.6                           |     | x   |             |
| QM17     | 2         | 15                  | 3.5                    | x     | 0.5                           |     | x   |             |
| QM18     | 1         | 15                  | 3.5                    | x     | 1.0                           |     | x   |             |
| QM19     | 3         | 15                  | 3.5                    | x     |                               |     |      |             |
| QM20     | 1         | 15                  | 4.5                    | x     |                               |     |      |             |
| QM21     | 1         | 15                  | 4.8                    | x     |                               |     |      |             |
| QM22     | 1         | 15                  | 5.0                    | x     |                               |     |      |             |
| QM23     | 3         | 10                  | 3.5                    | x     | 0.2                           |     | x   | x         |
| QM24     | 2         | 10                  | 3.5                    | x     |                               |     |      |             |
| QM25     | 1         | 10                  | 3.5                    | x     |                               |     |      |             |
| QM27     | 3         | 10                  | 3.5                    | x     | 0.2                           |     | x   | x         |
| GS02$^a$| 4         |                     | x                      | 0.5   |                               |     | x   |             |
| GS03$^a$| 3         |                     | x                      | 1.0   |                               |     | x   |             |
| QMP3$^b$| 3         |                     | x                      | xb    |                               |     |      |             |
| QMP4$^b$| 8         |                     | x                      | xb    |                               |     |      |             |

Note. Pore pressure gradient $\Delta P_f$ is given in the column for permeability.  
$^a$Samples and measurements from Aben et al. (2016), porosity measured through a different approach.  
$^b$Samples and measurements from Aben et al. (2017b), permeability obtained from Darcy flow with constant downstream pressure and fluctuating flow rate and upstream pressure.

The laboratory was kept at a constant temperature of 20 °C during measurements. The mean pore fluid pressure $P_f$ was set to 5 MPa for all measurements. The pressure gradients varied depending on the approximate permeability of the sample and are listed in Table 1. The hydrostatic confining pressure $P_c$ was increased by steps of 5 MPa from 10 MPa up to around 35–40 MPa and then decreased with a similar step size down to 10 MPa. Permeability and $P$ wave arrival times were measured at each hydrostatic pressure step. Specific storage capacity $S$ was obtained during the confining pressure steps in between permeability measurements. During such a step, one pore fluid reservoir was disconnected from the sample, whereas the second remained connected and at constant fluid pressure. This allowed us to monitor the displaced pore fluid volume from the system as the effective pressure ($P_{\text{eff}} = P_c - P_f$) was changed. The amount of pore fluid expelled from the sample $\Delta V_{pf}$ was obtained by correcting for the volume change of the sample holder setup (e.g., high-pressure tubing and pore fluid distribution plates. See Text S2 for details). From the last few pressure steps (25–35 MPa loading and unloading), the drained compressibility of the rock $\beta_d$ was calculated:

$$\beta_d = \frac{V_{0}^{\text{total}} - \Delta V_{pf} - \Delta V_{\text{solid}}}{V_{0}^{\text{total}}} \frac{1}{\Delta P_{\text{eff}}},$$

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Figure 3. (a) Peak stress of high strain rate loading tests on initially intact samples versus maximum strain rate (open squares) and yield strain rate (small blue squares). Upper and lower error bar markers represent the input and output stress, respectively. Same peak stress range applies to yield strain rate symbols. UCS was taken from Aben et al. (2016). (b) Dissipated energy versus residual axial strain for samples subjected to single (squares) and repeated (circles) loadings. Dashed lines show how dissipated energy is accumulated at each repeated loading for individual samples. Upper and lower error bar markers are dissipated energies calculated with input stress and output stress, respectively. Pulverization strain threshold is similar to Aben et al. (2016).

where $V_{total}^0$ is the initial total volume of the sample, and the volume change of the solid skeleton of the rock is given by

$$\Delta V_{solid} = V_{solid}^0 \beta_s P_{eff},$$

(2)

where $V_{solid}^0$ is the initial volume of the solid skeleton and $\beta_s$ is the compressibility of the solid skeleton. $\beta_s$ is set to $2 \times 10^{-11}$ Pa$^{-1}$, which is a representative value for crystalline rock (Brace, 1965). Nonetheless, the term $\Delta V_{solid}$ is several orders of magnitude smaller than $\Delta V_{pf}$. The specific storage capacity is computed as (Wibberley, 2002; Wibberley & Shimamoto, 2005)

$$S = (\beta_f - \beta_s) + \phi(\beta_f - \beta_s),$$

(3)

where $\beta_f$ is the compressibility of the pore fluid.

Fracture densities were derived from X-ray $\mu$CT scans of a number of samples (Table 1, scanned at the 3-SR and SIMAP laboratories in Grenoble, France) at a voxel size of 12 $\mu$m (spatial resolution of 24 $\mu$m), except for Samples QMP3 and QMP4 that were scanned at a 15 $\mu$m (spatial resolution of 30 $\mu$m). The fracture network was obtained by manually tracing fractures in three to four slices per sample. The slices were oriented perpendicular to the sample axis and spread equidistant along the length of the sample. The fracture density was given by the cumulative length of the fracture network normalized by the intact surface area of the sample, so that fracture opening is not taken into account. The spatial resolution of the CT scans only allows for tracing of large aperture fractures and therefore an underestimation of the cumulative fracture length compared to fracture tracing on optical microscope or SEM images. However, the frailty of the samples and some residue of grease and water residue after mechanical loading and permeability measurement prevented production of high-quality thin sections to do so. All seven X-ray $\mu$CT data sets have similar resolution and therefore yield internally consistent fracture density data. The large aperture fractures that could be traced also contribute most to the overall permeability of the sample.

3. Results

3.1. Mechanical Data

Postmortem samples were defined as either fractured or pulverized, where mechanical pulverization is recognized from the presence of a second strain rate peak during which the sample loses all strength (Aben et al., 2016)—thus losing all cohesion and load-bearing capacity under uniaxial conditions at the sample scale. This was further confirmed by a visual check for the presence of a pulverized texture (large amount of non-localized fracture damage and small fragment sizes). We note that the amount of fracture damage varies in pulverized material as defined here. This has also been observed in the field (Rempe et al., 2013) and in the
Figure 4. (a) Porosity increases with increasing total dissipated energy, for both single (squares) and successive (circles) high strain rate loadings. Pulverized samples (with the most dissipated energy) do not show a clear break in the trend. (b) Porosity increases as a function of residual axial strain. The increase is stronger for single high strain rate loadings compared to successive high strain rate loadings.

laboratory (Ghaffari et al., 2019) and seems to be controlled by strain rate (Ghaffari et al., 2019). Dynamically fractured samples are load bearing and cohesive during the loading tests and include an elastic unloading interval in their stress-strain curves. Visually, these samples show distributed main fractures that traverse several grains.

The successively loaded samples progressively weakened, as indicated by the decrease in peak stress, the decrease in Young's modulus, and the increase in permanent strain and dissipated energy (Figure 2b). The peak stress of the rock transcends the quasi-static uniaxial strength (measured by Aben et al., 2016) and...
Figure 6. Permeability at 30 MPa effective pressure (5–10 MPa effective pressure for the lowest permeability sample) versus yield strain rate (top) and total dissipated energy (bottom) for samples subjected to single (squares) and repeated (circles) loadings. For successively loaded samples, the yield strain rate of the first load has been taken.

Permeability increases with higher peak strain rate and higher yield strain rate (Figure 3a). Note that the quality of stress equilibrium deteriorates at higher strain rates as evidenced by the larger gap between input and output stresses, especially for successively loaded samples near the pulverization threshold. A large amount of energy was dissipated to achieve a certain amount of axial strain at low strain; near the pulverization strain threshold, the dissipated energy per strain unit strongly decreases (Figure 3b). This trend is similar for both single and successive loadings. Such a break in slope is typically observed near the pulverization strain threshold (Aben et al., 2016; Doan & Billi, 2011; Fondriest et al., 2017), where the sample no longer has a load-bearing capacity.

3.2. Porosity

Porosity initially increases by a few percent only up to a dissipated energy of 1 MJ m\(^{-3}\). At higher dissipated energy, porosity increases more steeply with dissipated energy up to 20% (Figure 4a). Again, single loadings and successive loadings follow the same trend. Pulverized samples (with the highest amount of energy dissipated, Figure 4) do not deviate from the trend set by fractured samples. The porosity of single loaded samples is higher relative to the porosity of successively loaded samples for the same amount of residual axial strain (Figure 4b).

3.3. Permeability

Permeabilities across the whole range of damage intensities vary from \(\kappa < 10^{-20}\) to \(\kappa \approx 10^{-13}\) m\(^2\) (Figure 5). We observe a distinct difference between fractured samples and pulverized samples, regardless of single or successive loadings. The permeability of fractured samples decreases nearly log linearly with increasing effective pressure (Figure 5). The permeability of pulverized samples does not show a pressure dependence at lower effective pressures. It increases at several measurements (asterisks in Figure 5), which we infer are due to some small shear offsets and the removal of sample material washed out by the fluid flow, which is supported by the observation of fine particles in the pore fluids after the measurements. During hydrostatic unloading, the permeability of dynamically fractured samples shows some hysteresis (Figure 5, small symbols). The permeability of the pulverized samples behaves more erratically during unloading (Figure 5). A second hydrostatic pressure cycle has been performed on Sample QM27, which shows a more stable permeability versus pressure relation. These results indicate that pulverized rock has initial high-permeability values that decrease when subjected to pressure cycles.

Permeability increases with yield strain rate for samples subjected to a single loading (Figure 6). Pinpointing a representative strain rate for samples subjected to repeated loadings is more difficult because the input stress wave does not apply a similar strain rate history on intact and damaged samples due to (1) lowered elastic properties of the damaged sample and (2) the increased impedance mismatch causes a larger part of the stress wave to be reflected at the input bar-sample interface. Hence, we regard the yield strain rate of the first loading best suited for comparison with the samples subjected to a single loading. Permeability (measured after the series of repeated loadings) increases with the yield strain rate of the first loading, although this strain rate dependency is less than for single loadings (Figure 6). The dependence of permeability on dissipated energy instead of strain rate better reflects the effect of the entire deformation history over multiple loadings. Here we assume that all dissipated energy creates new
fractures. This is justified by a strain-dissipated energy relation which is the same for single and repeated loadings (Figure 3b, Aben et al., 2016). Permeability of samples subjected to single loadings increases log linear with dissipated energy (Figure 6). Samples subjected to repeated loadings show a larger scatter, and a clear trend cannot be inferred (Figure 6). Permeability of pulverized samples—either subjected to single or successive loadings—is similar at $\kappa \approx 10^{-14}$ m$^2$.

Bernabé et al. (2003) proposed that the relation between permeability and porosity contains information on the process of permeability enhancement or reduction with increasing confining pressure in a single sample. For the least deformed sample (QM12), we find that porosity and permeability both decrease with increasing confining pressure (Figure 7). With increasing confining pressure on samples subjected to more dynamic fracture damage (QM16 and QM18), we see a stronger permeability drop relative to the drop in porosity (Figure 7). We see a similar trend for pulverized Sample QM27, although the drop in permeability over a 30 MPa confining pressure range is less than for the dynamically fractured samples. The other pulverized samples reveal a reverse trend with increasing confining pressure: Permeability increases strongly with a minor decrease in porosity (Figure 7). Note that only data measured during confining pressure loading is shown. The change in power law dependence

![Figure 8](image1.png)  
**Figure 8.** Storage capacity versus permeability at 30 MPa effective pressure.

![Figure 9](image2.png)  
**Figure 9.** (a) X-ray $\mu$CT slice perpendicular to the sample axis of Sample QM10 (pulverized, single loading) and the traced fracture network below. (b) X-ray $\mu$CT slice perpendicular to the sample axis of Sample QM23 (pulverized, successive loadings) and the traced fracture network below. (c) High aperture fracture density versus permeability. Error bars indicate the standard deviation from the mean. (d) High aperture fracture density versus dissipated energy.
between permeability and porosity in a set of samples subjected to progressive deformation can also reveal changes in the deformation process (Bernabé et al., 2003). Our data suggests that a power law coefficient $\alpha$ increases from approximately $\alpha = 1-2$ at low porosity (relatively undamaged samples) to $\alpha = 8-10$ at higher porosity (Figure 7).

### 3.4. Storage Capacity

Storage capacity was measured on six samples with higher permeabilities ($\kappa = 10^{-18} - 10^{-14}$ m$^2$) and was defined on the highest pressure steps of the hydrostatic unloading interval (see Text S2 for more detail). Storage capacities vary from $7.3 \times 10^{-11}$ Pa$^{-1}$ for fractured rock to $2.7 \times 10^{-10}$ Pa$^{-1}$ for pulverized rock (Figure 8).

### 3.5. Microstructures

The X-ray $\mu$CT slices reveal a structure typical for dynamically fractured and pulverized rocks recovered from laboratory experiments where zones of very fine material alternate with zones of apparently intact rock (Figures 9a and 9b). The zones of fine-grained material could not be fully traced at the resolution of the X-ray $\mu$CT. The permeability increases with an increase in high aperture fracture density below 2 mm/mm$^2$. At higher fracture density, the permeability remains more or less stable (Figure 9c). Fracture density increases linearly with dissipated energy, also at higher fracture densities obtained from pulverized samples (Figure 9d). This reflects the nonlinear dependence of permeability with the microstructure of the rock.

### 4. Discussion

#### 4.1. Hydraulic Characteristics of Dynamic Fracture Damage

The porosity increases up to 20% (Figure 4), which is uncharacteristic for fractured rock. Postmortem loss of small particles and rotation of fragments may be responsible for the opening of large pores. For pulverization under confining pressure, we expect the largest porosity values to be somewhat lower. The porosity-dissipated energy relation is the same for single and successive loaded samples (Figure 4a) but not for equal amounts of (cumulative) residual axial strain (Figure 4b). For the same amount of strain, successively loaded samples show a lower amount of porosity than single loaded samples. We suggest that successively loaded samples accommodate relatively more strain by sliding along preexisting fractures formed during preceding loadings, which does not create additional porosity. During single loadings on the other hand, strain is accommodated relatively more by creation of new fractures, thereby creating porosity, rather than by sliding on preexisting fractures as these are simply not yet formed. This implies a larger amount of microfractures for single loaded samples relative to successively loaded samples for the same amount of dissipated energy, which is confirmed by the fracture density counts of Samples QM10 (single loaded) and QM23 and QM27 (successive loaded): QM10 has a higher fracture density, while dissipated energies are similar to the successively loaded samples (Figure 9d).

Permeability hysteresis during a hydrostatic pressure cycle is commonly observed (Behnsen & Faulkner, 2011; Pérez-Flores et al., 2017; Wang et al., 2016). Here, the dynamically fractured samples show such behavior (Figure 5). In contrast, the erratic permeability behavior observed in pulverized samples during a first hydrostatic pressure cycle indicates that the internal damage structure is subject to additional permanent change, aside from standard structural changes that cause common hysteresis. The small particle sizes in pulverized rocks (Barber & Griffith, 2017; Muto et al., 2015) may affect the permeability; fluid flow causes erosion of small particles, which may be deposited elsewhere in the pulverized rock-forming geopetal-particle-aggregates (Schröckenfuchs et al., 2015). Here, we have found these particles in the pore fluid system, and the X-ray $\mu$CT images reveal that some larger aperture fractures are filled with fine-grained material that was potentially transported from elsewhere in the sample. For two of the pulverized samples, the porosity-permeability relation indicates that a small reduction in porosity increases permeability (Figure 7); this is in agreement with internal rearrangement of connectivity between pore spaces to enhance permeability (Bernabé et al., 2003). Porous rock such as sandstone is known to collapse at high hydrostatic pressure (e.g., Wong & Baud, 2012), and a similar type of local pore collapse may also occur in pulverized rock. Also, minor amount of slip along a main fracture has been shown to cause a significant increase in permeability (Pérez-Flores et al., 2017) and is likely to have occurred in our samples during the increase in confining pressure. The permeability measured on pulverized samples is thus not representative for a stable damage structure but reflects a transient permeability. A first-order approximation of a stable permeability in pulverized rock is given by the second confining pressure cycle of Sample QM27, which is 1 order of
We observe the increase in connectivity qualitatively in the X-ray until the sample behaves more as a granular aggregate. A progressive deformation during dynamic loading from dilatant microcracking to pervasive microcracking, explained by reported values of $\alpha$ (Bernabé et al., 2003). The change in power law exponent may be $\alpha$ increases with increasing porosity (Bernabé et al., 2003). The change in power law exponent may be $\alpha$ increases with increasing porosity (Bernabé et al., 2003): $\alpha \approx 1–2$ is expected for dilatant microcracking at an advanced stage, where a well-connected microcrack network has formed. For more diffuse microcracks, such as expected from thermal cracking, $\alpha$ increases to 4–5. We see an even stronger increase in $\alpha$ between dynamically fractured and pulverized rocks, similar to that reported for cataclastic compaction in low-porosity (<15%) sandstones (Bernabé et al., 2003). The change in $\alpha$ thus suggests a progressive deformation during dynamic loading from dilatant microcracking to pervasive microcracking, until the sample behaves more as a granular aggregate.

Permeability increases as a function of increasing damage (dissipated energy, Figure 6) and as a function of strain rate (Figure 6). However, a clear trend cannot be quantified. Between samples that experienced varying degrees of damage, the exponent $\alpha$ of the power law relation between permeability and porosity changes from $\alpha \approx 1–2$ to $\alpha \approx 8–10$ (Figure 7). This means that the relative connectivity of microfractures increases with increasing porosity (Bernabé et al., 2003). The change in power law exponent may be explained by reported values of $\alpha$ on low-porosity crystalline rock (Bernabé et al., 2003): $\alpha \approx 1–2$ is expected for dilatant microcracking at an advanced stage, where a well-connected microcrack network has formed. For more diffuse microcracks, such as expected from thermal cracking, $\alpha$ increases to 4–5. We see an even stronger increase in $\alpha$ between dynamically fractured and pulverized rocks, similar to that reported for cataclastic compaction in low-porosity (<15%) sandstones (Bernabé et al., 2003). The change in $\alpha$ thus suggests a progressive deformation during dynamic loading from dilatant microcracking to pervasive microcracking, until the sample behaves more as a granular aggregate.

We observe the increase in connectivity qualitatively in the X-ray $\mu$CT fracture networks. A progressive increase in connectivity of microfractures, the connective regime, is expected for low-porosity crystalline rock subjected to quasi-static deformation rates (Guéguen & Schubnel, 2003; Zhu & Wong, 1999). A modeling effort by Perol and Bhat (2016) for dynamic fracturing in an initially intact rock shows a similar trend. Our results confirm this permeability evolution with increasing dynamic fracture damage. The highest permeability is reached once all microfractures have coalesced in the connected regime. Beyond this point, the rock is pulverized and a fracture-based model becomes inadequate—the rock has effectively become a granular aggregate (Perol & Bhat, 2016). In this granular regime, we measure a permeability that remains stable with increasing amounts of damage (Figure 9c).

Rock samples subjected to high strain rates and short loading durations are either dynamically fractured or pulverized. The differences between both type of damage have been reported previously in terms of mechanical properties (Aben et al., 2016; Doan & Gary, 2009; Yuan et al., 2011), microstructures (Barber & Griffith, 2017), and acoustic velocities (Aben et al., 2016). Here, we show that the stabilized permeability of pulverized samples and highly fractured samples is within the same order of magnitude. A distinct transition in nominal porosity from fractured to pulverized samples is not observed. Hence, the fluid flow properties from dynamic fracturing to pulverization follow a continuous trend and do not show a similar threshold or transition as observed in mechanical data. Fluid flow properties of single-event pulverized rocks and pulverized rock from successive loadings are similar and do not provide a means to distinguish the two. Nonetheless, successively loaded samples may accommodate more cumulative strain along preexisting microfractures without increasing permeability, as evidenced by the porosity evolution with strain (Figure 4b).

### 4.2. Implications for Fault Zones

In section 4.2.1, we discuss how the damage zone permeability profile prior to a rupture (Figure 10a, profile $\kappa_1$) changes during (profile $\kappa_2$) and after a single earthquake rupture (profile $\kappa_3$) and how the damage zone magnitude lower than the initial transient permeability measured during the first hydrostatic pressure cycle

(\text{from } \kappa = 5 \times 10^{-13} \text{ to } 5 \times 10^{-14} \text{ m}^2, \text{ Figure 5}).

High strain rate damage was introduced at uniaxial conditions, which are representative for loading at the Earth surface. The permeability values were obtained at effective pressures up to 31 MPa, equalling roughly 2 km depth for hydrostatic fluid pressure. The pulverization microstructures produced by uniaxial compressive loading are anisotropic, based on acoustic velocity measurements (Aben et al., 2016). Acoustic measurements on pulverized field samples from the San Andreas fault show similar anisotropy (Rempe et al., 2013), and these samples have been exhumed from at most 1–2 km (Chester, 1999; Dor et al., 2006). This suggests that uniaxial pulverized microstructures do not differ significantly from shallow pulverized samples, and the permeability values are representative for shallow pulverization. Moreover, permeability values measured on pulverized rock sampled from the San Jacinto fault zone in California (Morton et al., 2012) are within the same order of magnitude as those measured here. Alternatively, transient permeability measured on pulverized samples during the first confining pressure cycle may represent permeability after a high strain rate pulverization event under tensile loading conditions, after which the stress field is restored—equivalent to an increase in confining pressure during our measurements. However, the microstructure of fragmented rock in radial isotropic tension and pulverized rock in compression may not be similar at similar strain rates (Griffith et al., 2018).

Rock samples subjected to high strain rates and short loading durations are either dynamically fractured or pulverized. The differences between both type of damage have been reported previously in terms of mechanical properties (Aben et al., 2016; Doan & Gary, 2009; Yuan et al., 2011), microstructures (Barber & Griffith, 2017), and acoustic velocities (Aben et al., 2016). Here, we show that the stabilized permeability of pulverized samples and highly fractured samples is within the same order of magnitude. A distinct transition in nominal porosity from fractured to pulverized samples is not observed. Hence, the fluid flow properties from dynamic fracturing to pulverization follow a continuous trend and do not show a similar threshold or transition as observed in mechanical data. Fluid flow properties of single-event pulverized rocks and pulverized rock from successive loadings are similar and do not provide a means to distinguish the two. Nonetheless, successively loaded samples may accommodate more cumulative strain along preexisting microfractures without increasing permeability, as evidenced by the porosity evolution with strain (Figure 4b).
Figure 10. (a) Schematic map view of a Mode II shear rupture. Rupture direction is from right to left, and shear stress decreases from peak stress at the rupture tip down to residual frictional strength at the end of the slip-weakening zone $R$, shown below the map view. Breakdown work $W_b$ is dissipated along the slip-weakening zone (b), whereas fracture energy $\Gamma$ is dissipated within the rupture tip process zone (c). Part of the fracture energy and breakdown work is dissipated as dynamic fracture damage in the wall rock; the width of the zone depends mostly on rupture velocity. Given the right circumstances, thermal pressurization (TP) is active at some distance behind the rupture tip process zone. Pore fluid pressure diffuses from the pressurized fault core into damaged wall rock further down the slip-weakening zone. Note that relative distances depicted here are arbitrary. Permeability evolution during rupture is shown by three fault-perpendicular profiles (small inset): $k_1$ on the locked part of the fault, $k_2$ behind the rupture tip process zone, and $k_3$ behind the total dynamic shear stress drop. (b) Schematic shear stress versus slip diagram depicting the total dynamic shear stress drop during rupture, including definitions for slip-weakening distance $\delta_0$ and breakdown work $W_b$. Small inset: see panel (c). (c) Schematic shear stress versus slip diagram showing the transition from dissipation of only fracture energy $\Gamma$ to energy dissipation by other processes contributing to $W_b$. Shape of shear stress versus slip curve based on quasi-static failure experiment in intact granite (Aben et al., 2019).

Permeability profile would evolve over successive earthquakes. These results will illustrate that pulverization near the fault core can occur close to the rupture tip (Figure 10a). Hence, pulverized rock may be in place prior to the activation of some slip-weakening mechanisms, such as thermal pressurization (Figure 10a). In section 4.2.2, we discuss how thermal pressurization is impacted by high-permeability damage zone rock.

We do this by presenting simple estimations of the hydraulic diffusivity and thermal pressurization factor in the damage zone wall rock, which are key parameters in assessing the damping effect on thermal pressurization (Figure 10a).

4.2.1. Fault Damage Zone Structure for Fluid Flow
First, some discussion is warranted on the origin of the off-fault strain rates responsible for coseismic dynamic fracturing. Off-fault strain rates arise from the velocity at which the rupture propagates and from the amplitude of the off-fault stresses around the rupture tip. The amplitude depends on the dynamic shear stress drop $\Delta \tau$ from its peak value at the rupture tip down to a residual shear stress on the slipping part.
of fault. The shear stress on the fault drops over a certain fault length, called the slip-weakening zone \( R \) (Figure 10a). The amount of slip accumulated at \( R \) is the slip-weakening distance \( \delta_0 \) (Figure 10b). The shear stress in the slip-weakening zone is often approximated as a linear decrease down to a certain distance behind the rupture tip (Figure 10a) (Ida, 1972; Palmer & Rice, 1973) or is estimated by assuming a certain slip-weakening mechanism, such as flash heating (Brantut & Viesca, 2017) or thermal pressurization (Viesca & Garagash, 2015). For simplicity, we adopt a linear slip-weakening law, where \( R \propto \delta_0/\Delta \tau \). A shorter slip-weakening distance, for a given constant stress drop, increases off-fault stresses and gives rise to a stress singularity if \( \delta_0 = 0 \). Note that the rise in dynamic off-fault stresses occurs in advance of the rupture tip as well as behind it (Figure 10a).

The breakdown work \( W_b \) invested to reduce the shear stress to the residual frictional strength is given by the area underneath the shear stress versus slip curve down to the residual shear stress (Figure 10b). \( W_b \) can be determined from seismological data (Kanamori & Rivera, 2006; Tinti et al., 2005), where \( W_b \) varies from \( 10^2 \)–\( 10^8 \) J m\(^{-2} \) as a function of total coseismic slip (Nielsen et al., 2016). For linear slip weakening, \( W_b = 0.5\delta_0\Delta \tau \), and thus, a higher breakdown work for a constant stress drop increases \( \delta_0 \) and reduces the off-fault stresses amplitude.

It is important to realize that \( W_b \) is an umbrella term for various dissipative processes that contribute to lowering the shear stress on the fault. Part of \( W_b \) is the shear fracture energy \( \Gamma \) dissipated to propagate the rupture by a unit length (Freund, 1990). \( \Gamma \) is dissipated within a process zone surrounding the rupture tip (Figure 10a) and is of the order of \( 10^4 \) J m\(^{-2} \) for intact crystalline low-porosity rock at upper crustal conditions (Aben et al., 2019; Lockner et al., 1991; Wong, 1982, 1986). Both \( W_b \) and \( \Gamma \) are dissipated on and off the fault as latent heat and new fracture surface area, including off-fault dynamic fracture damage (Figure 10a). Although fracture energy \( \Gamma \) is always dissipated earliest during rupture (i.e., closest to the rupture tip, Figure 10c), dissipation of \( \Gamma \) may overlap with some other constituent dissipative processes of \( W_b \). Constituent dissipative processes contributing to \( W_b \) that partly overlap or follow in the wake of the rupture tip process zone may be affected by dynamic off-fault damage and underlying changes in hydraulic properties resulting from dissipation of \( \Gamma \) in the rupture tip process zone itself. We will discuss one such constitutive dissipative process, thermal pressurization, in section 4.2.2.

We now establish the change in hydraulic properties in the damage zone due to dynamic fracturing for two scenarios: (1) The fault experiences the entire dynamic shear stress drop down to the residual frictional strength (i.e., \( W_b \) has been dissipated, Figure 10b) and (2) only the rupture tip process zone has passed by (i.e., \( \Gamma \) has been dissipated, Figure 10c). The resulting permeability profiles will represent schematic profiles \( k_1 \) and \( k_3 \) in Figure 10a. Scenario 2 is an end-member scenario drafted on the assumption that \( \Gamma \) is dissipated prior to other slip-weakening mechanisms. From quasi-static shear failure experiments in intact rock (Aben et al., 2019; Lockner et al., 1991; Wong, 1982, 1986), it is known that \( \Gamma \) is dissipated prior to a sharp change in slope as further weakening of the fault commences (Figure 10c). The existence of such a break in slope “separating” dissipation of \( \Gamma \) and \( W_b \) has also been shown for dynamic shear rupture following a rate-and-state weakening law rupture simulation (Barras et al., 2020). The advantage of Scenario 2 is that an a priori knowledge on further slip-weakening processes is not needed—after all, the effect of rupture-induced dynamic off-fault damage on these slip-weakening processes is yet unclear. The disadvantage is that the residual shear stress after the stress drop from dissipation of \( \Gamma \) must be kept constant for the linear elastic fracture mechanics (LEFM) model that we use here, whereas it would actually drop further by other slip-weakening mechanisms.

We use the same LEFM model for Scenarios 1 and 2 described above but with different parameter values that are detailed after the model description. Our idealized fault is embedded in intact host rock and subjected to shear rupture at constant velocity in the \( x \) direction. The fault ruptures as a Mode II shear crack at sub-Rayleigh wave speed so that the stress field formed around the rupture tip is described by linear elastic fracture mechanics. LEFM solutions for a singular stress field (Freund, 1979, 1990) have been used previously to estimate the distance from the interface at which pulverized rocks are encountered (Aben et al., 2017a; Doan & Gary, 2009; Griffith et al., 2018; Reches & Dewers, 2005). Following this solution, peak strain rate decays by \( y^{-1.5} \), where \( y \) is fault perpendicular distance. Here, we eliminate the stress and strain rate singularities by adopting the nonsingular slip-weakening LEFM solution of Poliakov et al. (2002). The shear stress along the fault drops from its initial value at the rupture tip to its residual value at \( R \) (Figure 10a).
The rate of decrease in hydraulic properties is expressed in its quasi-static limit as (Rice, 1980):

$$R = \frac{9\pi}{16(1-\nu)} \frac{\mu W_b}{\Delta \tau^2}, \quad (4)$$

where $\nu$ is Poisson’s ratio and $\mu$ is the shear modulus. $R$ at higher rupture velocity is a function of $R_0$ and the rupture velocity, given by Poliakov et al. (2002). The nonsingular solution for the stress field is given by equations (A3) and (A11) in Poliakov et al. (2002). We consider strain rate as the key parameter for dynamic fracturing. Stress is ignored for now, as the transient rupture-related stress components need to be added to the tectonic background stress, which varies per fault. The peak volumetric strain rate with distance $y$ was obtained from the rupture tip stress field and rupture velocity, similar to Doan and Gary (2009) and Reches and Dewers (2005). We adopt the elastic properties and density of intact quartz-monzonite used in our experiments. The rupture velocity is set at 0.9 of the Rayleigh wave speed of the material.

For Scenario 1, we obtain strain rate fields for a rupture with $W_b = 0.5$ MJ m$^{-2}$ ($\Delta \tau = 100$ MPa and $\delta_0 = 0.01$ m) and a rupture where $W_b = 5$ MJ m$^{-2}$ ($\Delta \tau = 100$ MPa and $\delta_0 = 0.1$ m). These ruptures have slip-weakening zones of 2.6 and 26 m, respectively.

Off-fault strain rates in Scenario 2 were calculated for ruptures with constant $R_0 = 0.13$ m and with $\Delta \tau = 100$, 10, and 1 MPa so that $\delta_0 = 0.5$, 0.05, and 0.005 mm, respectively. The rupture with $\Delta \tau = 100$ MPa is representative for intact crystalline rock, where $\Gamma = 25$ kJ m$^2$ (Aben et al., 2019). Fracture energies for the two other ruptures are 1 or 2 orders of magnitude lower and may be representative of weaker fault rocks.

We obtained fault-perpendicular permeability values by comparing the yield strain rates that the samples experienced in the experiments with the peak volumetric strain rates (Figures 11a and 11c). We simulated successive ruptures in Scenario 1 simply by converting the yield strain rate of the first loading of a successive loadings series to distance.

For the ruptures modeled in both scenarios, peak strain rate decays by $r^{-1.5}$ at larger normalized distance from the rupture, following the asymptotic solution by Freund (1990) (Figures 11a and 11c). The peak strain rate flattens off closer to the rupture tip. The inflection point occurs at 10% of the fault normalized distance, but the absolute strain rate values are larger for smaller $R_0$ or higher stress drop. We also see that for some chosen parameters ($R_0 = 26$ m in Figure 11a), the pulverization threshold for successive loading of quartz-monzonite (100 s$^{-1}$) (Aben et al., 2016) is not surpassed and pulverization in compression is not expected.

After a single earthquake where $R_0 = 2.6$ m, the damage zone permeability profile perpendicular to the fault rises sharply at 14 cm distance from the fault (Figure 11b). The permeability does not increase once the rocks are pulverized closer to the fault at about 5 cm distance. Although measurement results are uncertain on pulverized rocks (see section 4.1), they suggest that the permeability remains stable. Hence, there is a maximum permeability limit for fault damage zone rocks (Figure 11b). Successive ruptures (two or three in the case of our experiments) extend the band of pulverized rock outward as the pulverization threshold for predamaged rock drops. The permeability profile migrates outward as well, and the slope of the profile decreases (Figure 11b). This decrease arises from the stronger increase in permeability with progressive dynamic fracture damage below the pulverization threshold relative to the permeability increase for progressive damaging of already pulverized rock. For further loadings, we expect the slope to decrease even more. As argued by Aben et al. (2016), the zone of pulverized rock will eventually stabilize at a set distance from the fault plane. This distance is a anchor point for the highest permeability in the profile. Pulverization in the rupture tip process zone is predicted for all the chosen parameters in Scenario 2, although pulverization is limited to near-fault distances within 0.65 mm (for $\Delta \tau = 1$ MPa) to 68 mm (for $\Delta \tau = 100$ MPa) from the fault plane (Figure 10c). Permeability in the rupture tip process zone is expected to increase strongly at fault-parallel distances of 1.3 cm or less from the fault plane, for $\Delta \tau = 10$ MPa (Figure 10d).

Pulverization occurs somewhere in the slip-weakening zone $R$. For Scenario 1, $R$ is of the order of meters to tens of meters. We have not fully explored where in the slip-weakening zone the conditions are met for pulverization and dynamic fracture damage in terms of strain rate and stress. This may be meters away from the rupture tip (given that $R$ is tens of meters). In this case, the off-fault dynamic fracture damage and change in hydraulic properties may not have a large effect on slip-weakening mechanisms, that is, it occurs...
Figure 11. (a) Peak volumetric strain rate as a function of normalized distance from the fault plane, based on the elastodynamic stress field around a nonsingular slip-weakening rupture at 90% of the Rayleigh wave speed (solutions by Poliakov et al., 2002). Results shown are for ruptures accommodating a total dynamic stress drop (Figure 10b). Δτ = 100 MPa with \( R_0 = 2.6 \) m (top curve in (a)). (b) Permeability versus normalized distance from the fault plane for single loadings (squares) and successive loadings (circles), for a rupture with \( R_0 = 2.6 \) m (top curve in (a)). These results are similar to schematic permeability profile \( \nu_3 \) in Figure 10a. Permeabilities were measured at 31 MPa effective pressure. (c) Peak volumetric strain rate as a function of normalized distance from the fault plane in the rupture tip process zone (Figure 10c), for Δτ = 10, 10, and 1 MPa so that \( \delta_0 = 0.5, 0.05, \) and 0.005 mm. The dashed line indicates the strain rate pulverization threshold in compression, and the maximum distances for pulverization are indicated. (d) Permeability versus normalized distance from the fault plane for single loadings, for the rupture with Δτ = 10 MPa. These results are similar to schematic permeability profile \( \nu_2 \) in Figure 10a.

“too late” in the rupture. As discussed earlier, \( W_b \) measured for a real earthquake may not be dissipated following a linear slip-weakening law. Scenario 2 provides the other extreme: a steep drop in shear stress near the rupture tip. Here, pulverization occurs within 0.13 m from the rupture tip and will therefore most likely have occurred before or during the activation of other slip-weakening processes (Figure 10a). Both Scenarios 1 and 2 are somewhat simplified representations of stress drops along faults. The location at which the right set of strain rate and stress conditions is met for pulverization relative to the rupture tip remains a subject for future research.

We note that we used a pulverization criterion for compression to predict fault-perpendicular distances at which we expect dynamic fracturing and pulverization. Loading in compression occurs on one side of the fault, whereas tensile loading is anticipated on the opposite side of the fault. Pulverization in radially isotropic tension (Griffith et al., 2018) or volumetric tension (Xu & Ben-zion, 2017) will decrease the strain.
rates necessary for dynamic fracturing and pulverization. This results in a larger fault-perpendicular distance at which pulverization is expected to occur. Given the same material on both sides of the fault, we therefore predict a larger zone of pulverized rock, and with it a broader damage zone permeability profile, on the side of the fault subjected to tensile loading. Rupture simulations that allow for dynamic off-fault damage indeed confirm this (Thomas & Bhat, 2018).

**4.2.2. Slip-Weakening Mechanisms: Thermal Pressurization**

Frictional heating during fast fault slip causes rapid thermal expansion of solid fault material and, when present, pore fluids. Pore fluid pressure increases due to the compressibility contrast between solid (low compressibility) and pore fluid (high compressibility). Consequently, the effective normal stress on the fault is reduced—that is, the fault weakens with slip. This mechanism is called thermal pressurization and accommodates slip at temperatures lower than those expected for a dry fault (Lachenbruch, 1980). Thermal pressurization acts in the principal slip layer of fault, which is typically a layer of fault gouge up to 1 mm in width. At a small amount of slip (i.e., directly in the wake of the rupture tip), the gouge layer is considered to be in undrained conditions, and fluids in the gouge layer do not yet communicate with those in the damage zone (Rice, 2006; Rempel & Rice, 2006) (Figure 10a). The principal slip zone is under drained conditions at prolonged slip, and pore fluid overpressure in the slip zone diffuses into bounding, stationary gouge (Rice, 2006; Rempel & Rice, 2006), or into the damage zone wall rock (Figure 10a).

The material that bounds the slipping gouge zone thus affects thermal pressurization only during prolonged slip at drained conditions. This may occur in two ways: (1) The initial pore pressure within the damage zone is reset to a lower value by dilation. Depending on how fast the fluid pressure reequilibrates, this may affect the diffusion and flow of overpressurized pore fluids from the slip zone to the damage zone at prolonged slip. (2) In addition to the decreased pore fluid pressure, the changed hydraulic properties of the pulverized damage zone rock increase pore fluid pressure diffusion and flow from the slip zone (Figure 10a). To analyze these two effects in detail, prior knowledge on the fault zone structure is crucial: The transition zone from gouge in the principal slip zone to pulverized rock adjacent may contain nonslipping bounding gouges and cataclasites that form impermeable barriers between the pressurized slip zone and pulverized rock. Other unknowns are the initial fluid pressure and the absolute dilation-induced fluid pressure drop during rupture. As these parameters and the fault zone geometry are fault specific and may change within a fault zone, we do not aim for a full numerical solution. Instead, we aim for a first-order estimate of how thermal pressurization is affected by near-fault pulverization.

To do so, we calculate the path-averaged thermal pressurization factor and the hydraulic diffusivity (following the approach of Brantut & Platt, 2017), which together give an estimate for the efficiency of thermal pressurization. We use the mechanical and hydraulic properties measured on three end-member materials: experimentally sheared gouge with a granitic mineral content (Zhang et al., 1999), natural clay-bearing gouge from the Median Tectonic Line in Japan (Wibberley, 2002; Wibberley & Shimamoto, 2003), and experimentally pulverized rock in this study. We consider three scenarios (Figure 12a): Case 1 involves slip in a body of pulverized rock. Case 2 encompasses slip in granitic gouge, surrounded by pulverized rock that will experience the same pressure and temperature path of that in the gouge (i.e., the pulverized rock is instantaneously heated and pressurized). This setup reflects a “fresh” fault in crystalline rock. In Case 3, a clay-bearing gouge layer is bound by pulverized rock, because the fault core of a preexisting fault hosted in crystalline rock is expected to be enriched in clay content (Rowe & Griffith, 2015). The pulverized rock will again experience the same pressure and temperature path of the gouge.

We use the classic framework to describe thermal pressurization: In the slipping fault gouge layer, pressure $P_f$ and temperature $T$ evolve with time $t$ as (Lachenbruch, 1980; Rice, 2006; Rempel & Rice, 2006):

\[
\frac{\partial P_f}{\partial t} = \Lambda \left( \frac{\gamma f}{\rho c} + \alpha_{th} \frac{\partial^2 T}{\partial y^2} \right) + \alpha_{hy} \frac{\partial^2 P_f}{\partial y^2},
\]

with $y$ as the distance parallel to the fault, $r$ as the shear stress on the fault, $\gamma$ as the slip rate, $\rho c$ the heat capacity, and $\alpha_{th}$ the thermal diffusivity. $\Lambda$ is the thermal pressurization factor:

\[
\Lambda = \frac{\lambda_f - \lambda_n}{\beta_f - \beta_n},
\]

with $\lambda_f$ and $\lambda_n$ the thermal expansivity and $\beta_f$ and $\beta_n$ the compressibility of the pore fluid and pore space, respectively. We assume that $\lambda_n$ is the same as the thermal expansion of solid grains in granite ($2.4 \times 10^{-5}$...
Figure 12. (a) Three cases considered in illustrating the impact of hydraulic properties of pulverized rock on thermal pressurization. For Cases 2 and 3, the pulverized wall rock will experience the same pressure and temperature path of the gouge. (b, d, f) Thermal pressurization factor $\Lambda$ versus depth in pulverized rock (green curves) and in granitic or clay-rich gouge (black curves) for the three cases shown in (a). The thermal pressurization factor is shown for the adiabatic, undrained limit (early slip, thick curves) and for the slip-on-a-plane limit (prolonged slip, thin curves). (c, e, g) Hydraulic diffusivity $\alpha_{hy}$ versus depth in pulverized rock (green curves) and granitic or clay-rich gouge (black curves) for the three cases shown in (a). Hydraulic diffusivity only exists for prolonged slip.

$c^{-1}$, Rice, 2006). $\alpha_{hy}$ is the hydraulic diffusivity:

$$\alpha_{hy} = \frac{\kappa}{\phi(\beta_n + \beta_f)\eta},$$

where $\kappa$ is permeability, $\eta$ is pore fluid viscosity, and $\phi$ is porosity. The term $\phi(\beta_n + \beta_f)$ is the storage capacity $S$.

We consider two limits in terms of slip: early slip and prolonged slip. For early slip, the hydraulic diffusivity term does not exist as thermal pressurization occurs at undrained adiabatic conditions in a slip zone of finite width. At prolonged slip, the width of the principal slip zone vanishes and it is considered as a single plane. Analytical solutions for these cases are valid only for constant material properties and slip velocity (Lachenbruch, 1980; Rice, 2006). In reality, the material properties of both the solid and fluid vary as a function of fluid pressure and temperature. We therefore compute $P_f-T$ path-averaged values for all material properties described above (Rice, 2006). The expected range of temperature and fluid pressure with slip is predicted by using the nominal material properties (at preslip pressure and temperature). This $P_f-T$ range is then used to calculated path-averaged values (for more details, see Brantut & Platt, 2017). For Cases 2 and 3, we obtain the $P_f-T$ range for slip in gouge and apply them to pulverized rock as well.
Depth- and temperature-dependent water properties were taken from WATER95 (Junglas, 2009). The depth range of our analysis is between 2 and 10 km, since large-scale pulverization is a shallow crustal phenomenon. The slip rate is set to 1 m s$^{-1}$ and the coefficient of friction to 0.6. The pressure dependence of permeability for pulverized rock is taken from the unloading interval of the permeability-effective pressure curve to eliminate most of the hysteresis (Figure 5). To do so, we fitted an exponential function to the unloading permeabilities of pulverized Samples QM10, QM23, and QM27 and took the average of the fits, so that $\kappa = 2 \times 10^{-14}(-2.8 \times 10^{-8}P_{\text{eff}})$. The same was done to obtain the change in $\phi$ with effective pressure (equations (1) and (2)), so that $\Delta \phi = 0.124(-1.8 \times 10^{-9}P_{\text{eff}})$. For the two gouge materials, we used the pressure dependence of permeability and porosity given by Brantut and Platt (2017), who derived them from measurements by Wiibberley (2002), Wiibberley and Shimamoto (2003), and Zhang et al. (1999).

For early slip in pulverized rock (Case 1), $\Lambda$ varies between 0.7 and 0.85 MPa K$^{-1}$, whereas for prolonged slip $\Lambda$ drops to about 0 MPa K$^{-1}$ (Figure 12b), and peak temperatures exceed 1000 °C over the entire depth range. Over the entire depth range, the hydraulic diffusivity is of the order of 10$^{2}$ mm$^2$ s$^{-1}$ (Figure 12c), which is similar to the nominal hydraulic diffusivity prior to slip.

Early slip in fresh granitic gouge (Case 2) yields a thermal pressurization factor $\Lambda$ that varies roughly between 0.25 and 0.7 MPa K$^{-1}$ (Figure 12d). $\Lambda$ in the bounding pulverized rock varies between 0.85 and 0.94 MPa K$^{-1}$. At prolonged slip, the thermal pressurization factor for pulverized rock drops by about 0.2 MPa K$^{-1}$ over the entire depth range but remains higher than that of gouge. Temperatures reach 1000 °C at around 9 km depth. The hydraulic diffusivity of pulverized rock surrounding slipping gouge is more than 2 orders of magnitude higher than that of the gouge (Figure 12e) and about 1 order of magnitude higher than that predicted for pulverized rock in Case 1.

For Case 3, we find that $\Lambda$ in the slipping clay-rich gouge varies between 0.6 and 0.8 MPa K$^{-1}$ during early and prolonged slip both (Figure 12f). For early slip, $\Lambda$ in the bounding pulverized rock is similar to that found for Case 2 and only slightly higher than $\Lambda$ in the clay-rich gouge. For prolonged slip, $\Lambda$ in pulverized rock increases significantly, with values ranging from 1.1 MPa K$^{-1}$ at 10 km depth to 1.8 MPa K$^{-1}$ at 2 km depth. The hydraulic diffusivity of the clay-rich gouge is between 10$^{-2}$ and 10$^{-1}$ mm$^2$ s$^{-1}$ (Figure 12g)—the lowest of the fault materials studied here. In contrast, the hydraulic diffusivity of the pulverized wall-rock is the highest at 10$^{3}$–10$^{4}$ mm$^2$ s$^{-1}$.

The estimate for thermal pressurization during slip in pulverized rock (Case 1) shows that pulverized rock is initially more effectively pressurized than granitic or clay-rich gouges. At prolonged slip however, thermal pressurization is switched off and the rock experiences frictional melting. It is however unrealistic to assume that material properties in pulverized rock remain unchanged when sheared—they would most likely evolve to fine cataclasite or gouge-like properties by comminution and display hydraulic properties as measured by Boutareaud et al. (2008) on fine cataclasite rock. Nonetheless, we do show that for very high temperatures, the thermal pressurization factor becomes effectively zero in pulverized rock.

The parameter $\Lambda$ for pulverized wall rock (Cases 2 and 3) only affects thermal diffusion (second term in brackets, equation (4)). As $\Lambda$ is higher for pulverized rock than for both gouges at prolonged slip, the pore fluid pressure in pulverized wall rock rises with an increase in temperature. However, the hydraulic diffusivity is extremely high in pulverized rock; thus, pore fluid pressure diffuses quickly near the boundary of the slipping gouge layer. Moreover, we suggest that a more effective thermal pressurization processes in the gouge (i.e., clay-rich gouge in Case 3, which has somewhat higher $\Lambda$ but lower $\eta_{\text{eff}}$ compared to granitic gouge) result in an order-of-magnitude increase in hydraulic diffusivity in pulverized wall rock. This is supported by Case 1, where thermal pressurization is nonexistent and the hydraulic diffusivity of pulverized rock is similar to its nominal value. Thus, the hydraulic diffusivity of the pulverized wall rock may reduce thermal pressurization effectiveness as the pressurized slipping zone decreases in width at intermediate slip, which may lead to local frictional melt of the principal slip zone (Brantut & Mitchell, 2018).

It should be noted that we study pressure bleeding from a pressurized fault core based on measurements made on the sample scale. Larger fractures and secondary fault strands may act as pore pressure sinks on the larger scale. This is evidenced by fluidized granular flow that resulted in fault gouge injection in wall rock fractures and other dilational structures (Kirkpatrick & Shipton, 2009; Rowe et al., 2012). Along-fault heterogeneity in material and the rupture of several fault strands in a single event may also prevent widespread thermal pressurization (Boutareaud et al., 2008).
Our first-order estimates of how thermal pressurization is affected by pulverization are not entirely conclusive: The thermal pressurization factor of pulverized rock is slightly higher to those in the slipping gouge zone, but they may be offset by orders-of-magnitude increase in hydraulic diffusivity as a function of thermal pressurization effectiveness. Nonetheless, these results do highlight the importance of understanding the impact of the hydraulic diffusivity of pulverized rock on pore fluid pressure diffusion.

5. Conclusion
In this study, we measured the fluid flow properties of low-porosity crystalline rock subjected to high strain rate uniaxial compression experiments. Single and successive loading experiments were performed to study the effect of progressive accumulation of high strain rate fracture damage. We show that porosity and fracture density increase as a function of dissipated energy. Permeability increases 6 orders of magnitude ($k < 10^{-20}$ to $k \approx 10^{-13}$ m$^2$) with increasing strain rate and with increasing dissipated energy, and porosity increases by 15%. We argue that permeability of pulverized samples does not increase relative to the most fractured samples just below the pulverization threshold. The fluid flow properties of pulverized samples subjected to single or successive loadings are similar. The permeability evolution with porosity is described by a power law exponent of $\alpha \approx 1 - 2$ to $\alpha > 8$. The obtained fluid flow properties give an estimate of the permeability profile through a fault damage zone resulting from the total dynamic stress drop during an earthquake. We expect a steep permeability increase when approaching the fault core, up to the pulverization boundary where we expect permeability to remain stable or to decrease. We argue that successive earthquake events reduce the slope of the permeability profile, and the permeability profile migrates outward toward a steady-state distance. We also suggest that pulverization occurs near the rupture tip in a narrow zone near the fault plane. This may affect slip-weakening mechanisms following in the wake of the rupture. A first-order assessment of thermal pressurization as a slip-weakening mechanism in the presence of pulverized rock has been conducted. These estimates suggest that pulverized wall rock is effective in damping thermal pressurization at prolonged slip.

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