Sample dilation and fracture in response to high pore fluid pressure and strain rate in quartz-rich sandstone and siltstone

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1 Natural hydraulic fractures (NHFs) are inferred to form where pore pressure exceeds the least compressive stress by an amount equal to the tensile strength of the rock. We improved upon an experimental protocol that meets the NHF criterion within cylindrical samples with the most tensile effective stress parallel to the sample axis. The effective tensile stresses achieved during these experiments ranged from 17–47 MPa. The pore fluids used had higher viscosities than water and the axial strain rate was rapid ($\sim 10^{-3}$ s$^{-1}$) to delay dissipation of fluid pressure by flow. Four experiments on St. Peter Sandstone samples and two on an Abo Formation siltstone sample were performed under these conditions and under drained conditions. None of the drained experiments resulted in failure, but all of the sandstone and one of the siltstone samples fractured in response to elevated pore pressures. Consistent with field and theoretical studies, mechanical heterogeneity was a first order control on fracture location. In the absence of mesoscopic heterogeneity, fracture location coincided with the maximum pore pressure. Samples responded to elevated pore pressures and differential stresses by dilating, the magnitude of which was sufficient to achieve atmospheric pore pressure. Samples failed 2 to 250 s after experiencing the greatest pore pressures, when the effective stresses were no longer tensile. Thus, the high pore pressures and effective tensile stresses experienced early in the experiments were sufficient to fracture the rocks, even though they were not sustained until failure. These results provide insight into processes of fluid-driven fracture formation.


1. Introduction

Elevated fluid pressures are proposed to create effective tensile stresses large enough to fracture rock even under far-field compressive stress conditions [Secor, 1965]. Natural hydraulic fractures (NHFs) are inferred to form where pore fluid pressure exceeds the least compressive stress by an amount equal to the intrinsic tensile strength of the rock. However, the micro-mechanical processes that lead to failure as well as the strain rate dependence of these processes are not well understood. Further, although it has been demonstrated theoretically and with field examples that fractures initiate at mechanical heterogeneities, the size and character of heterogeneity required to localize fractures has not been extensively studied in the field or laboratory [Pollard and Aydin, 1988; Lacazette and Engelder, 1992]. Experimental studies of fluid-driven fractures have primarily focused on induced hydraulic fractures [Haimson and Fairhurst, 1967], which form by hydrologic and mechanical processes distinct from those that produce NHFs [Renshaw and Harvey, 1994]. Induced hydraulic fractures form in response to the high-pressure injection of fluids, forcing a fracture to initiate at, and propagate from, the injection location. Further, the fluid pressure in the crack is maintained by an external source, which forces crack propagation. In contrast, NHFs form in response to the elevation of pore fluid pressure within a rock volume; their location is presumed to be largely controlled by mechanical heterogeneities. The fluid pressure in the crack may decrease as the fracture grows if the rock permeability is low, which would cause the fracture to arrest until the pressure in the crack has increased sufficiently [Renshaw and Harvey, 1994]. Therefore, the primary difference between NHFs and induced hydraulic fractures is that propagation of NHFs is limited by the hydrologic properties of the rock. Because they have not been studied experimentally, the current understanding of NHF initiation and propagation is based on a large body of theoretical work and numerical modeling [Secor, 1965; Rice and Cleary, 1976; Pollard and Holzhausen, 1979; Lacazette and

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Engelder, 1992; Renshaw and Harvey, 1994; Miller, 1995; Boutt et al., 2009]. Models for formation include continuum mechanics models of poroelastic deformation and linear elastic micro-mechanical models. There is no existing theoretical or empirical model for the effects of strain rate and dilatancy on NHF formation in porous rocks. Field studies of fractures rely heavily on theory to interpret stress and pore pressure conditions at failure [Pollard and Aydin, 1988; Savalli and Engelder, 2005]. However, field studies suffer from the limitations that the stresses and pore pressure at failure must be inferred from structures that have often experienced complex histories. Therefore, these studies rarely allow us to test theoretical mechanical failure criteria.

[3] We improved upon an experimental protocol described by Boutt et al. [2009] that produces both a pore fluid pressure gradient parallel to the long axis of a solid cylindrical sample and effective stresses high enough to overcome the tensile strength of the rock, allowing us to initiate NHFs in the laboratory. These experiments are distinct from the rupture test, which produces an induced hydraulic fracture in an internally pressurized hollow cylindrical sample, because water is not being forced into the propagating fracture and the fracture location is not determined by the experimental set-up. We report experimental results from a suite of sandstone and siltstone samples with variably developed mechanical anisotropy and heterogeneity, which allowed us to evaluate the roles of mechanical and hydrologic heterogeneity in fracture localization. In addition, we compare the strain response of each sample to the behavior predicted by two-dimensional models, parameterized with measured sample properties, that couple poroelastic deformation and fluid diffusion. We show that dilatant material behavior deviated from linear elastic, reducing pore pressure to atmospheric prior to failure; however, the samples still formed mesoscopic fractures.

[4] Four experiments were performed on samples from the St. Peter Sandstone and two on samples from the Abo Formation siltstone. The St. Peter Sandstone samples are fine to medium-grained quartz arenites with spatially variable amounts of cement and primary structures. The Abo Formation siltstone is a fluvial quartz-rich siltstone with low permeability and mm-scale bedding laminae defined by grain shape preferred orientation and oxide, dolomite, and clay abundance. Experiments were performed on both a mesoscopically homogeneous sample and cores exhibiting varying degrees of heterogeneity and anisotropy, which were variably oriented with respect to applied stresses. The results of this study demonstrate that although sample-scale mechanical heterogeneity and anisotropy affect the location of fracture initiation, neither is required for fractures to nucleate and grow.

[5] Section 2 provides relevant theoretical context for our study, section 3 provides a description of the experimental protocol, and section 4 provides sample descriptions, all of which are necessary to understanding the experimental results (section 5).

2. Theoretical Background

2.1. Fracture Nucleation and Localization

[6] The stress conditions leading to formation of fluid-driven fractures are generally studied by assuming rock deformation follows Biot’s constitutive equations of poroelasticity prior to failure and that deformation is coupled to fluid flow using Darcy’s Law [Biot, 1940; Pollard and Aydin, 1988; Detournay and Cheng, 1991]. In the simplest model, fractures form when far-field effective stresses overcome the characteristic tensile strength, $T$, of the rock. Employing the convention that tensile stress is negative, this is expressed as the effective stress criterion as

$$\sigma'_I = \sigma_{ij} - \delta_{ij} p \leq -T,$$

where $\sigma_{ij}'$ is the effective stress, $\sigma_{ij}$ is the total stress, and $\delta_{ij}$ is the Kronecker delta function, and $p$ is the pore pressure.

[7] Field evidence, experimental evidence, and theoretical work indicate that fractures, including NHFs, initiate at mechanical flaws [Griffith, 1920; Tapponnier and Brace, 1976; Pollard and Aydin, 1988; Lacazette and Engelder, 1992]. In detail, failure is localized by elevated stresses at the boundaries of inclusions, microcracks, and pores. Crack tips localize tensile stresses, such that the magnitude of the tensile stress is dependent on the aspect ratio of the crack [Griffith, 1920]. This localized tension leads to though-going fractures initiating on favorably oriented elongate cavities, including elongate pore spaces and micro-cracks, which grow and coalesce into macroscopic fractures. Thus, the grain-scale structure of the rock influences fracture initiation and propagation.

[8] The stress concentration at a crack tip (or elongate pore) is thus a function of the shape and size of the crack as well as the far-field stresses. In this context, the stress intensity factor, $K_I$, of an isolated crack can be expressed as

$$K_I = \sigma Y \sqrt{\pi c},$$

where $\sigma$ is the far field stress, $c$ is the crack half-length, and $Y$ is a geometric factor. For penny-shaped cracks, $Y = 1.13$ [Atkinson and Meredith, 1987]. The theory of linear elastic fracture mechanics defines a critical stress-intensity factor, $K_{IC}$, such that a crack will propagate once the characteristic $K_{IC}$ of the rock is reached. For geologic materials, the assumption that a crack-like flaw must exist for a fracture to propagate is generally met due to common elongate pores and/or mesoscale cracks. However, in high-porosity rocks, the stress concentrations at pore boundaries interact and cannot be easily approximated. In this case, the effective stress criterion (1) may be more useful in predicting failure.

[9] At steady state conditions, the effect of pore pressure on $K_I$ is calculated using the effective stress, $\sigma'$, in place of $\sigma$ in (2) [Pollard and Holzhausen, 1979; Detournay and Cheng, 1991]. Detournay and Cheng [1991] showed that immediately following a change in pore pressure, rock deformation is governed by the undrained moduli everywhere but at the crack tip, which causes strength hardening. The stress intensity, $K_I$, is lower than equation (2) and equal to

$$K_I = \frac{1 - \nu_u}{1 - \nu} \sigma' Y \sqrt{\pi c},$$

where $\nu$ and $\nu_u$ are the drained and undrained Poisson’s ratios, respectively. In the long-term limit $K_I$ approaches equation (2). This analysis emphasizes the temporal evolution of the stress state surrounding a pressurized fracture.
during loading, but not growth of the fracture. During failure, however, the rock surrounding a crack may be inelastically damaged, rather than responding as a poroelastic medium. And under transient conditions, such as during crack propagation, the stress and pore pressure conditions that drive crack growth are less certain.

[10] Using a theoretical interpretation of their experimental data, Bruno and Nakagawa [1991] suggest that the local pore pressure contributes to the strain energy available for fracture propagation and that, on a larger scale, pore pressure gradients also drive fracture propagation. However, there is no consensus on either of these issues [Detournay and Boone, 1993; Bruno, 1994], in part because the nature and importance of poroelastic versus inelastic deformation during fracture growth have not been explored.

[11] Implicit in both the effective stress and critical-stress intensity criteria is the assumption that once a given criterion is met, fracture initiation occurs immediately and propagation is rapid [Secor, 1965; Atkinson and Meredith, 1987]. Rice and Simons [1976] showed that, when the material at mode II crack tips is assumed to behave elastically, slow strain rates (drained conditions) favor crack growth over undrained conditions. However, neither the time it takes a throughgoing mode I fracture to localize nor the localization mechanisms have been carefully measured or observed. Thus, we have no record of the elastic and inelastic deformation mechanisms that accompany NHF formation.

2.2. Dilatancy

[12] Dilatancy has been observed in rocks subjected to compressive stresses during experimental tests on both crystalline and sedimentary rocks [e.g., Handin et al., 1963; Brace et al., 1966]. In rocks subjected to high mean stress, dilatancy is accommodated by propping open microcracks, thereby reducing the pore pressure of the system and reducing the effective stress [Brace et al., 1966; Lockner and Stanchits, 2002]. As a result, an apparent strengthening occurs relative to that expected from the non-dilated system with higher pore fluid pressure. Brace et al. [1966] noted that when failure is attributed to elevated pore pressure, dilatancy must accompany deformation and should be included in the Mohr and Griffith failure criteria. However, we still have a very basic understanding of the coupled relationship between elevated pore fluid pressure and micro-mechanical deformation.

[13] Most analyses and experiments completed to date address shear deformation, because of its relevance to the earthquake nucleation process and the difficulty of producing tensile failure in the laboratory [Brace et al., 1966; Rice, 1975]. The existing experimental work shows that both high differential stresses and rapid deformation enhance dilatancy in porous and crystalline rocks under compressive stress states [Brace et al., 1966; Zoback and Byerlee, 1975; Schmitt and Zoback, 1992; Lockner and Stanchits, 2002].

[14] Schmitt and Zoback [1992] showed that dilatant strengthening occurs in crystalline rocks during rupture tests. With elevated pore pressures, samples show apparent stiffening and tensile strengthening compared to dry samples subjected to the same internal pressurization rate. This occurs because the effective stress is actually less than expected due to dilation in the high -pore pressure samples. These experiments do not quantify the amount of dilation nor do they investigate the micro-mechanics that lead to failure.

[15] Although deviations from elastic behavior leading to shear failure have been recognized and studied, models of NHF formation still assume that under tensile stress conditions rock deforms elastically until failure. The lack of experimental data on NHF formation has limited scientific understanding of the stress, strain, and pore fluid pressure conditions that lead to the formation of opening-mode fractures observed in the field.

3. Experimental Protocol

[16] In a traditional triaxial experiment, $\sigma_1$ is axial and it is impossible to increase the pore pressure above $\sigma_1$ without creating a gap (essentially a hydraulic fracture) between the sample and the endcap. We therefore refined an existing experimental protocol introduced by Boutil et al. [2009] that uses a triaxial setup to meet equation (1) while maintaining contact between the sample and the endcaps. Samples are cylindrical cores 12.5 cm long and 5 cm in diameter. The objective is to produce an effective tensile stress parallel to the core axis that meets the effective stress criterion (Figure 1). We use the notation that $a$ is the direction parallel to the core axis and $r$ is the radial direction, and refer to the imposed boundary stresses as $\sigma_a$ and $\sigma_r$ to distinguish between the applied stresses and the resulting principal stresses within the sample.

[17] Unlike most triaxial experiments, initial conditions are set with $\sigma_r > \sigma_a$. First, the confining pressure is increased to $\sigma_r$ and the axial load is simultaneously increased to $\sigma_{ai} = \sigma_r - 2$ MPa. Next a spatially uniform pore fluid pressure, $p$, is raised to $p_i = \sigma_r - 3$ MPa (Figures 1a and 1d). The goal of the procedure is to reduce the axial load to $\sigma_{af} = 2$ MPa as the pore pressure is simultaneously reduced to 0 MPa at the core ends. Mechanical equilibrium is reached before hydrologic equilibrium, because the mechanical equilibrium time constant is much smaller than the hydrologic equilibrium time constant. This results in $p > \sigma_{af}$ everywhere except immediately adjacent to the end caps and produces both a pore fluid pressure gradient, $\nabla p$, and an effective tensile stress $\sigma_1'$ parallel to the core axis (Figures 1b–1d). Under the appropriate initial conditions, this results in pore pressures and effective tensile stresses much greater than the tensile strength of the rock. In practice, because reduction of the pore pressure is achieved by manually opening the top and bottom fluid ports to atmospheric pressure, pore pressure is reduced first to maintain $\sigma_a > p$ at the sample ends and $\sigma_r$ is reduced ~1 s later. The axial strain rate was $\sim 10^{-3}$ s$^{-1}$ for these experiments and the data sampling rate was 4 Hz. If the effective stress criterion holds, we expect the rock to fracture immediately after the load is reduced. To test whether fractures formed during this process were fluid-driven or the result of the differential stress, a drained test was run on most samples prior to the NHF experiment. The same $\sigma_r$ and $\sigma_a$ used during the NHF experiment were used for the drained tests and the samples were saturated, but pore pressure was maintained at atmospheric pressure.

[18] Pore fluid viscosities greater than water were used in the experiments to decrease the diffusivity, $\kappa$, delay
dissipation of the pore pressure, allow adequate time to reduce the axial load, and maximize the likelihood of fracture initiation. The relationship between diffusivity and the experimental time constant is shown graphically in Figure 2. Pore fluids were chosen to achieve an elevated pore pressure near the core center for at least $\geq 3$ s; most samples have diffusivities resulting in elevated pore pressures for $>100$ s. Ethylene glycol was used in the Abo Formation experiments; oil-based Dow Corning ® fluids were used to saturate the St. Peter Sandstone samples. Samples 3_2 and 12_1 allowed them to maintain elevated pore fluid pressures for more than $100$ s. The 5_5 St. Peter samples maintained elevated fluid pressures for less than 10 s.

Experiments were conducted in the geomechanics facility at Sandia National Laboratories using a 1 MN load frame. Radial confining pressure was applied with Isopar ®, an isoparaffinic fluid solvent. Each sample was loaded between endcaps designed to allow maintenance of a differential stress. Felt metal distributor plates between the sample and the endcaps distributed the fluid pressure and allowed even fluid flow near sample ends. Each sample was jacketed with polyolefin shrink wrap and secured to the end caps with wire. The setup was modified and a strip of metal 9.60 mm wide and 0.10 mm thick was wrapped around the joint between the sample and the endcap to prevent

Figure 1. The procedure for producing NHFs in the laboratory. $\sigma_{ai}$ is the initial axial stress, $p_i$ is the initial pore pressure, $\sigma_{af}$ is the axial stress after reducing the end load, and $p_f$ is the pore pressure in the sample interior after reducing the end load. (a) Initially $\sigma_{ai}$, $\sigma_r$, and spatially uniform $p_i$ are applied such that $\sigma_{ai} = p_i + 1$ MPa and $\sigma_r = \sigma_{ai} + 2$ MPa. (b) $\sigma_{ai}$ and $p$ are simultaneously reduced at the core ends, which produces a pore pressure gradient parallel to the core axis, resulting in conditions that meet the failure criterion everywhere but at the sample ends. (c) Some time after $\sigma_{ai}$ and $p$ are reduced a fracture forms. The pore pressure near the fracture at the time of failure is primarily a function of the location of the fracture and diffusivity. (d) The stress state of each step (Figures 1a–1c). Black indicates stresses at the sample center (where $p$ gradually decreases over time) and gray indicates stresses at the sample ends (where $p = 0$). Following reduction of the axial load, differential stress increases until mechanical equilibrium is attained.

Following reduction of the axial load, differential stress increases until mechanical equilibrium is attained.

Table 1. The Experimental Stress and Pore Pressure Boundary Conditions for Each Core

<table>
<thead>
<tr>
<th>$\sigma_{ai}$ (MPa)</th>
<th>$\sigma_{af}$ (MPa)</th>
<th>$p_i$ (MPa)</th>
<th>$p_f$ (MPa)</th>
<th>$\mu$ (Pa s)</th>
<th>$\kappa$ (m$^2$/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.2</td>
<td>50</td>
<td>48</td>
<td>47</td>
<td>2</td>
<td>0.97</td>
</tr>
<tr>
<td>12.1</td>
<td>50</td>
<td>48</td>
<td>47</td>
<td>2</td>
<td>0.37</td>
</tr>
<tr>
<td>5.5 □</td>
<td>20</td>
<td>18</td>
<td>17</td>
<td>2</td>
<td>0.97</td>
</tr>
<tr>
<td>5.5</td>
<td></td>
<td></td>
<td>25</td>
<td>23</td>
<td>22</td>
</tr>
<tr>
<td>Abo ⊥</td>
<td>40</td>
<td>38</td>
<td>37</td>
<td>2</td>
<td>0.02</td>
</tr>
<tr>
<td>Abo</td>
<td></td>
<td></td>
<td></td>
<td>40</td>
<td>38</td>
</tr>
</tbody>
</table>

*Here $\sigma_{ai}$ is the initial axial stress, $p_i$ is the pore pressure, $\sigma_{af}$ is the axial stress after reducing the end load, $\mu$ is fluid viscosity, and $\kappa$ is sample diffusivity.
Cores prepared from the samples described in the text. (a–d) Cores from the St. Peter Sandstone; (e, f) cores from the Abo Formation siltstone. \( \perp \) designates cores oriented with long axes perpendicular to bedding; bedding is parallel to the axes of cores marked \( \parallel \). Note variable heterogeneity and anisotropy discussed in text.

the confining fluid from breaking the sample jacket for the last two samples tested, 3_2 and 12_1. During failed experiments on both of these samples, the sample jacket broke during unloading until the change was implemented. Axial strain was measured with two linear voltage differential transducers (LVDTs) attached to the endcaps and the radial strain was measured with two LVDTs attached to the sample with circumferential split rings. The samples were pre-saturated with the high viscosity fluids. Boundary loads for each experimental sample are provided in Table 1.

4. Samples

[20] We performed six NHF experiments on three sandstone samples (4 cores) and one siltstone sample (two cores) chosen to represent a range of hydrologic and mechanical properties. The sandstone samples were collected from the Ordovician St. Peter Sandstone [Mai and Dott, 1985]. Previous studies report a moderately well sorted mature sandstone with a grain composition that is \(~97\%\) quartz [Pitman et al., 1997; Kelly et al., 2007]. The St. Peter was chosen for these experiments because the sandstone is compositionally relatively simple; the variation in mechanical and hydrologic properties between and within samples is primarily controlled by variation in quartz overgrowth cementation with lesser variability in Fe-oxide and feldspar cementation, which simplified interpretation of experimental data.

[21] Sample 3_2 is the most homogeneous and isotropic sample studied (Figure 3a). Sample 5_5 has cm-scale bedding highlighted by variations in iron oxide cement and quartz cement abundance. The core 5_5 \( \perp \) was oriented with its axis perpendicular to bedding (Figure 3b) and core 5_5 \( \parallel \) with its axis parallel to bedding (Figure 3c). Sample 12_1, with cm-scale ellipsoidal to irregularly shaped concretions in which porosity is filled with iron oxide, allows consideration of the influence of distributed heterogeneity on fracture initiation in the absence of anisotropy.

[22] The Permian Abo Formation siltstone was chosen for these experiments because it contains more mineral phases than the St. Peter Sandstone and has well defined bedding laminae, which made it a suitable end-member sample. The mechanical and hydrologic characteristics of the Abo Formation siltstone were examined and published by Boutt et al. [2009]. The two cores from the Abo have mm-scale bedding defined by grain plus overgrowth preferred orientation and compositional variability, primarily variations in oxide and clay abundance. The samples tested are similar to the fine-grained fluvial quartzose siltstones described by Bensing et al. [2005], but were taken from coarser channel fill deposits. The core Abo \( \perp \) is oriented with its axis perpendicular to bedding and Abo \( \parallel \) has the core axis parallel to bedding (Figures 3e and 3f).

[23] Characterization of petrophysical and hydrologic properties from \( \mu \text{m} \) to cm scales was completed to evaluate the nature and scales of physical heterogeneity and/or anisotropy present within these samples. Bulk mechanical and hydrologic properties were measured using experimental cores or the equivalent and sections of cores. Destructive analyses were performed on samples taken within 25 cm laterally and within the same bedding interval as the cores used in the deformation experiments.

4.1. Petrophysical Characteristics

[24] To characterize samples at the grain and pore scale, thin sections were cut perpendicular to bedding. For samples 5_5 and the Abo siltstone, which have bedding visible at the hand sample scale, thin sections and analyses span multiple beds. Thin sections were imaged using a Hitachi S-3400 variable pressure scanning electron microscope at an accelerating voltage of 15 kA and an emission current of 60 \( \mu \text{A} \). Back-scattered electron (BSE) and cathode luminescence (CL) images were collected over the same area for each sample (Figures 4a and 4b). Composite images of St. Peter Sandstone, collected at 150x magnification, cover an area 3.70 by 2.25 mm. Images of the Abo siltstone were collected at 250x magnification to obtain similar statistics in the finer grained rock over an area 2.25 by 1.68 mm.

[25] Back-scattered electron mosaics were analyzed with the National Institute of Health (NIH) ImageJ software. The gray scale distribution and energy dispersive spectroscopy
analyses were used to determine the two-dimensional modal mineralogy and porosity of the samples (http://rsb.info.nih.gov/ij/) (Table 2). The BSE images and ImageJ software also were used to measure the size and distribution of pores in the samples. The largest pore size measured for each sample is reported in Table 2. Greyscale CL images were used to distinguish detrital quartz from cement in the St. Peter Sandstone samples. The proportion of cement was measured from tracings of quartz overgrowths using ImageJ (Table 2).

Figure 4
table 2. Modal mineralogy and grain/pore framework characteristics of experimental samples determined by analysis of bse images and cl images.

<table>
<thead>
<tr>
<th></th>
<th>detrital quartz (%)</th>
<th>detrital clay (%)</th>
<th>quartz cement (%)</th>
<th>fe oxide cement (%)</th>
<th>potassium feldspar cement (%)</th>
<th>carbonate cement (%)</th>
<th>mean grain diameter (mm)</th>
<th>max pore diameter (mm)</th>
<th>porosity (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.2</td>
<td>74.9</td>
<td>0.0</td>
<td>16.3</td>
<td>0.3</td>
<td>0.0</td>
<td>0.0</td>
<td>0.20</td>
<td>0.48</td>
<td>8.6</td>
</tr>
<tr>
<td>12.1 bulk</td>
<td>73.2</td>
<td>0.0</td>
<td>16.3</td>
<td>0.7</td>
<td>0.0</td>
<td>0.0</td>
<td>0.19</td>
<td>0.78</td>
<td>8.8</td>
</tr>
<tr>
<td>12.1 concretion</td>
<td>61.6</td>
<td>0.0</td>
<td>11.6</td>
<td>8.7</td>
<td>0.0</td>
<td>0.0</td>
<td>0.16</td>
<td>0.49</td>
<td>5.0</td>
</tr>
<tr>
<td>5.5 dark beds</td>
<td>73.5</td>
<td>0.0</td>
<td>2.8</td>
<td>2.5</td>
<td>4.5</td>
<td>0.0</td>
<td>0.20</td>
<td>1.30</td>
<td>15.8</td>
</tr>
<tr>
<td>5.5 light beds</td>
<td>73.8</td>
<td>0.0</td>
<td>2.8</td>
<td>2.5</td>
<td>4.5</td>
<td>0.0</td>
<td>0.20</td>
<td>1.30</td>
<td>22.8</td>
</tr>
<tr>
<td>Abo</td>
<td>70.6</td>
<td>14.8</td>
<td>n/a</td>
<td>0.5</td>
<td>0.0</td>
<td>0.0</td>
<td>7.0</td>
<td>n/a</td>
<td>2.40</td>
</tr>
</tbody>
</table>

*Clay phases are kaolinite and minor smectite and the carbonate is dominantly dolomite with minor calcite. The mean grain diameter includes quartz overgrowths (i.e. is the mechanically effective grain diameter). The maximum pore diameter in the Abo siltstone is the size of the largest microcrack observed.

[26] The detrital framework of all of the St. Peter Sandstone samples is 100% quartz with authigenic quartz overgrowths, minor authigenic potassium feldspar, and iron oxide cements in decreasing abundance. Pores are irregular to elongate in shape and occur with the long axis both sub-parallel and at high angles to bedding (Figures 4c and 4e). Heterogeneity in the samples is defined by variability in amount and spatial distribution of quartz and iron oxide cements.

[27] Scanning electron microscope images in Figures 4c–4g illustrate grain scale variability in these samples. Sample 3.2 is the most homogeneous and isotropic. It has a relatively uniform cement distribution, low oxide content, and no potassium feldspar overgrowths (Figure 4c and Table 2). Sample 12.1 is similar to 3.2 in bulk composition and cement abundance; however, it contains roughly ellipsoidal (though locally irregular) concretions 3–15 mm in the longest dimension with long axes commonly parallel to bedding, in which iron oxide largely fills the pore space (Figure 4d and Table 2). These areas have lower porosity and less quartz cement than the bulk rock, therefore we analyzed the bulk rock and concretions in separate image mosaics.

[28] Variations in magnitude of iron oxide cementation highlight bedding in sample 5.5. Iron oxide coats grains and forms grain-bridging contacts in dark beds (Figure 4f). Beds with higher iron oxide cement also contain more potassium feldspar, which occurs as overgrowths on quartz grains (Table 2). As a result of higher cement content, these beds have lower porosity (compare Figures 4e and 4f).

Image analysis on representative light and dark beds yielded porosities of 22.8% and 15.8%, respectively.

[29] The Abo Formation sample is a coarse-grained siltstone to very fine-grained sandstone dominated by detrital quartz, which we refer to as the Abo siltstone to emphasize the difference in average grain size between this and the St. Peter Sandstone. The siltstone contains distributed clay identified as detrital by Bensing et al. [2005]. An x-ray diffractometer was used to identify the clay mineralogy from a powdered bulk rock sample of the siltstone as primarily kaolinite with minor smectite. Cement is a combination of quartz overgrowths and carbonate (dolomite and minor calcite). Quartz overgrowths are ubiquitous whereas carbonate cement occurs in patches (Figures 4g and 4h). The small grain size made it impossible to separate the quartz overgrowths from original grains using CL images. Bedding in the Abo siltstone is defined by small variations in iron oxide content as well as quartz grain size. Locally there is a grain-shape preferred orientation; however, its relationship to bedding is spatially variable and ranges from subparallel to ~30° to bedding. Pores in the siltstone are numerous, but very small (Figure 4h). Intergranular microcracks oriented sub-parallel to sub-perpendicular to bedding laminae are locally evident in BSE images (Figure 4h).

The maximum microcrack length measured from the BSE mosaic is given as its maximum pore diameter in Table 2.

4.2. Mechanical Characteristics

[30] Two elastic parameters, bulk modulus (K) and Poisson’s ratio (ν), and tensile strength (T) were determined for each sample (Table 3). The bulk moduli were determined from the loading stage of each NHF experiment. Poisson’s ratios were measured by the Engineering department of the University of Massachusetts-Amherst on the experimental cores or equivalent cores. We estimated the tensile strength of the samples using Brazilian tests at the University of Wisconsin-Madison. Brazilian test specimens, disks 1.50 inches in diameter and 0.75 inches in thickness, were prepared and oriented to measure tensile strength perpendicular to the
Table 3. Results of Mechanical Tests, Flow-Through Permeability Measurements, and Porosimeter Porosity Measurements

<table>
<thead>
<tr>
<th>Sample</th>
<th>K (GPa)</th>
<th>μ</th>
<th>T (MPa)</th>
<th>k (m²)</th>
<th>n (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>3_2</td>
<td>5.2</td>
<td>0.26</td>
<td>1.7 ± 0.2</td>
<td>3 × 10⁻¹⁵</td>
<td>5.1</td>
</tr>
<tr>
<td>12_1</td>
<td>6.1</td>
<td>0.21</td>
<td>3.8 ± 0.2</td>
<td>1 × 10⁻¹⁵</td>
<td>4.1</td>
</tr>
<tr>
<td>5_5</td>
<td>1.6</td>
<td>0.22</td>
<td>0.5 ± 0.1</td>
<td>1 × 10⁻¹²</td>
<td>28.9</td>
</tr>
<tr>
<td>5_5</td>
<td></td>
<td></td>
<td>1.4</td>
<td>0.18</td>
<td>0.5 ± 0.1</td>
</tr>
<tr>
<td>Abo</td>
<td>11.0</td>
<td>0.19</td>
<td>6.8 ± 0.7</td>
<td>1 × 10⁻¹⁷</td>
<td>6.5</td>
</tr>
<tr>
<td>Abo</td>
<td></td>
<td></td>
<td>9.1</td>
<td>0.18</td>
<td>15.5 ± 0.9</td>
</tr>
</tbody>
</table>

*T is mean tensile strength, ±1σ is the standard deviation of three measurements of T, K is bulk modulus, μ is Poisson’s ratio, k is flow-through permeability, and n is the porosity determined using a helium porosimeter. Porosity was not measured on bedding parallel cores. 3_2 is the isotropic sample and 12_1 contains concretions.

The mechanical properties show variability that correlates well with the grain-scale structure of the samples. Samples with the highest bulk moduli and tensile strengths also have the lowest porosity (Tables 2 and 3). For the St. Peter Sandstone samples, tensile strength and bulk moduli increase with increasing quartz cement. Although 3_2 and 12_1, excluding concretions, have similar compositions and porosities (4–5%), 12_1 has both a higher bulk modulus (6.1 vs 5.2 GPa) and greater tensile strength (3.8 vs 1.7 MPa). The Abo siltstone is stronger in tension, both parallel (15.5 MPa) and perpendicular (6.8 MPa) to bedding, than any of the St. Peter Sandstone samples. The difference between bed-parallel and bed-perpendicular tensile strength in the Abo siltstone indicates a significant strength anisotropy. In contrast, bed-parallel and bed-perpendicular samples of 5_5 show no anisotropy in tensile strength. An anisotropy may exist, but is too weak to detect with a Brazilian test.

Figure 5. (a–f) Volume strain during drained and NHF (undrained) tests of St. Peter Sandstone cores and volume strain of the undrained tests of the Abo siltstone cores. Time is normalized by test duration, which is provided in the upper left-hand corner of each plot. The volume strain for the drained tests show both unloading (the actual drained test) and reloading of the sample to the initial experimental conditions. The undrained tests show much greater volume strain. Drained tests show little to no permanent strain.
4.3. Hydrologic Characteristics

The bulk permeability and porosity of the samples were determined by flow-through permeability and helium porosimeter measurements, respectively. The samples used for permeability measurements were cylindrical cores with the same dimensions as experimental cores. Two of the cores, 5_5 || and Abo ⊥, were the same as those used in the deformation experiments; the others were taken from adjacent areas of the same samples. Centimeter-scale porosity was measured using a helium porosimeter in the Engineering department of the University of Wisconsin-Madison on cylindrical cores 2.5 cm in diameter and ~5–6.5 cm long. Only bedding perpendicular cores were analyzed for samples with bedding (i.e., the 5_5 and Abo samples).

Flow-through permeability measurements for the suite of samples span 5 orders of magnitude (Table 3). Lower permeability correlates with more cement, lower porosity, higher tensile strength, and larger bulk modulus (Tables 2 and 3). The porosimeter provides lower values for porosity than image analysis does for samples 3_2 and 12_1 and higher values for 5_5 (Table 3). For 3_2 and 12_1, thin section images probably include unconnected pores not measured by the porosimeter. Additional reasons for the differences may include variability related to scale of observation.

5. Experimental Results

Drained tests were performed on all of the St. Peter Sandstone cores prior to each NHF experiment. The drained tests were performed by manually reducing the axial load to $\sigma_a = 2$ MPa at a strain rate similar to that used during the NHF test ($\approx 10^{-3}$ s$^{-1}$). Samples were held at these conditions for between 16 s (5_5 ||) and 80 s (3_2). The axial load and pore pressure were subsequently increased to the initial conditions of the NHF test. Unloading and reloading curves for the drained tests show negligible permanent volume change ($<0.01\%$), indicating inelastic deformation was not significant during the drained tests (Figure 5). Drained tests were not performed on Abo siltstone in our study. However, a similar sample of Abo siltstone was tested previously by Boutt et al. [2009]. In that case, the drained test did not produce a fracture in either the bedding perpendicular or
parallel cores, but the NHF test produced a fracture in the bedding perpendicular core.

Five of the six cores tested developed throughgoing fractures during the NHF experiments. Abo || was the only core that did not fail. The unloading paths for all of the cores show initial axial shortening as result of pore fluid draining from the system after the fluid ports were opened, followed by axial extension once the axial load was reduced (Figure 6). The axial strain rate during the initial stage of unloading was \( \frac{10^{-4}}{s} \). Sample extension dictated the axial strain rate required to maintain a 1 MPa axial load, which was \( \frac{10^{-4}}{s} \) for all samples. Throughgoing fractures developed after significant axial elongation, although the magnitude of elongation varied.

Samples 3_2, 12_1, and Abo \( \perp \) produced an audible sound indicating formation of a throughgoing fracture. Axial elongation and axial stress both increased synchronously with the sound, presumably as the radial load forced the fracture walls apart. The time to failure for these three samples varied from \( \sim 30 \) s to \( \sim 250 \) s after \( \sigma_a \) and \( p \) were reduced (Figure 6). We first attempted initial conditions of \( \sigma_r = 40 \) MPa, \( \sigma_{al} = 38 \) MPa and \( p = 37 \) MPa for these three samples, causing a fracture to form in Abo \( \perp \); however no fracture was produced in either 3_2 or 12_1. The initial conditions were increased to \( \sigma_r = 50 \) MPa, \( \sigma_{al} = 48 \) MPa and \( p = 47 \) MPa and fractures formed in both 3_2 and 12_1.

The two weakest, most permeable samples, 5_5 || and 5_5 \( \perp \), did not produce a sound at failure, nor did they exhibit the same increase in \( \sigma_a \) and \( \varepsilon_a \) the others did following failure. We interpret that failure of sample 5_5 \( \perp \) occurred at or prior to the sudden increase in axial strain (Figure 6c). Beyond this axial strain, no additional volume strain was observed. The increase in axial strain was most likely caused by deformation on surfaces of failure, which will be discussed later. The time to failure was between 1.25 s and 2.0 s; the exact timing is not obvious from the stress and elongation response. We infer that the sample failed by 8 s, because beyond this time the sample experienced no additional axial or volume strain.

The initial conditions for the test of 5_5 || were \( \sigma_{al} = 18 \) MPa and \( p = 17 \) MPa. A test of 5_5 \( \perp \) using the same conditions as 5_5 || did not result in failure. Subsequently, all initial conditions were increased by 5 MPa which resulted in failure. The axial extension rate was lower for 5_5 || than

Figure 7. (a–f) Volume strain during the NHF tests. For 3_2, 12_1, and 5_5 || (Figures 7a, 7b, and 7d) volumetric strain during unsuccessful tests (conducted at lower pore pressure) are shown. Horizontal lines indicate the volume strain required to reduce pore fluid pressure to 0 MPa for each experiment, calculated using equation (5). Note that in all cases, fractures formed only after significant core dilation, when pore fluid pressure was negligible. Cores experienced significantly less dilation during the unsuccessful tests.
5.5 \perp even though the initial pore pressure and effective tensile stresses were higher (Figure 6).

Volume strains experienced by the St. Peter Sandstone cores during drained tests are compared to volume strains during NHF tests to evaluate the effect of elevated pore pressures on deformation (Figure 5). During both drained and undrained tests the samples dilated. However, for all cores the volume strain response was much greater at elevated pore pressures. Volume strain was calculated from the average measured axial and radial strains. Strain was, however, probably heterogeneous prior to failure, due to spatial variation in pore fluid pressure, as well as at failure, due to the localization process. Radial strain is particularly sensitive to spatial variability; therefore, we also calculated the volume strain using each radial strain gauge alone and saw no significant difference between the different calculations of volume strains. However, the spatial variation in volume strain during the localization process is not known.

Presumably, both the dilation of pore space recorded by this volume sample and flow out of the sample contributed to a decline in pore fluid pressure over the course of the NHF experiments. The change in fluid pressure resulting from dilation, independent of the change due to fluid flow, is given by

$$\Delta p = \frac{(\epsilon_a + 2\epsilon_r)}{n \times c_f},$$

where $\epsilon_a + 2\epsilon_r$ is the sample volume strain calculated from the measured axial and radial strains, $n$ is porosity, and $c_f$ is the fluid compressibility. Therefore, the total dilation required to reduce pore pressure to zero (without taking into account flow out of the sample) is

$$\epsilon_a + 2\epsilon_r = p_0 \times n \times c_f.$$
5.5|| at the lower initial pore pressures that did not produce failure also are shown in Figures 7a, 7b, and 7d respectively.

6. Fracture Locations and Character

[41] Fracture locations are shown on images of the experimental cores in Figure 8. The most homogeneous sample, 3_2, developed one throughgoing fracture at the core center (Figure 8a). Sample 12_1 developed one throughgoing fracture and a series of non-throughgoing fractures, many of which intersect the edges of concretions. Abo ⊥ failed on a plane subparallel to bedding (Figure 8c). During sample saturation, the primary lamina on which the fracture formed bubbled more vigorously than the rest of the sample, suggesting it was more permeable than the rest of the core. This higher permeability likely reflects a higher porosity, weaker bed.

[42] The zones of failure in 5.5 cores were distinct from those in other samples, but similar to one another (Figures 8d and 8e). Both samples formed sets of tabular failure zones near the core ends with small acute angles to the core axis. In 5.5 ⊥, these features formed at small angles to bedding, whereas in 5.5 ||, they are oriented at high angles to bedding. The geometry and offset along the zones of failure in these samples provide evidence of shear displacement; significant sand production within the zones is evidence of disaggregation and/or cataclasis. The cross-cutting structures in these samples created ‘wedges’ that are displaced toward the center of the core relative to the adjacent pieces of the fractured sample (Figures 8d and 8e). The discrete lateral displacement of these wedge-shaped pieces was recorded during the experiment as an increase in axial elongation. These observations collectively suggest that the structures formed in these cores are mixed-mode deformation bands, which accommodated both extension and shear.

7. Poroelastic Models

[43] Two dimensional, plane strain, poroelastic models were created to evaluate the core responses to experimental conditions. Tensile failure criteria equations (1) and (2) both assume that failure occurs once the relevant criterion is met and it is generally assumed that deformation prior to failure is elastic. The models allow us to track spatial variations in stress, strain, and pore pressure up to the point of failure. We subsequently reconsider model assumptions and results in discussing the experimental results.

[44] Poroelastic models couple fluid diffusion and strain of the solid framework using Darcy’s Law and Biot’s constitutive equations of poroelasticity [Biot, 1940]. Data presented in section 5 were used to parameterize the models. Because the scaled heterogeneity of samples 12_1 and Abo is too fine to easily model, these samples were simplified as isotropic (Figure 9a). Heterogeneity is explicitly incorporated in models of the 5.5 samples (Figures 9b and 9c). To approximate the effects of anisotropy and elastic discontinuity on the poroelastic response, samples 5.5 ⊥ and 5.5 || were modeled with a bimodal distribution of beds with distinct elastic and hydrologic properties. The distribution of the two types of bedding is approximated as the “light” and “dark” bedding observed in hand sample. We estimated the elastic (Young’s modulus, E) and hydrologic properties (permeability, k) of each bedding type from the bulk elastic and hydrologic properties of the bedding-parallel and bedding-perpendicular cores and the volume of light and dark beds. The permeabilities of light and dark beds were estimated using cm-scale air-minipermeameter measurements [French, 2009].

[45] The material constants in the governing equations are Young’s modulus, E, Poisson’s ratio, ν, Lame’s parameters, G and λ, specific storage, S, hydraulic diffusivity, κ, and porosity, n. The values of the elastic parameters E, G, and lambda were calculated from ν and bulk modulus, K, measured during experiments. The specific storage was approximated by assuming that fluid compressibilities were negligible. The errors produced by this approximation are small because fluid bulk moduli are generally 3 orders of magnitude greater than sample bulk moduli, which are summed in the denominator of the storage equation. The porosities used in most models are helium porosimeter values, which we believe more accurately reflect the 3-dimensional hydrologic character of pores. For 5.5, we relied on image analysis of light and dark bedding to capture spatial variations in porosity.

[46] At time t = 0 s, a normal stress of σ, is applied in the x2 direction to the model sides, a normal stress of σai is applied in the x1 direction to the model ends, and p is increased to the initial pressure used in the experiment (Table 1). These conditions are held for t = 3000 s to ensure system equilibrium. At t = 3000 s, the pore pressure is reduced to p = 0 MPa at the model ends and σ, is incrementally decreased to σ, ≈ 2 MPa. Axial stresses in the models were reduced at the same rates as during the experiments.

[47] The models predict that the tensile stress exceeded the tensile strength immediately after the axial load was reduced and throughout the experiments. The maximum effective tensile stress anywhere within the sample prior to failure and the maximum tensile stress experienced near the failure plane are provided for each sample in Table 4. The effective tensile stress parallel to the core axis is greatest immediately after the end load is reduced. The stress intensity factors of the cores at this time were estimated from equation (2).
assuming that the poroelastic models provide a reasonable estimate of $\sigma_3'$ early in the experiments. In calculating $K_I$ from equation (2), we used the largest pore diameter or microcrack measured during image analysis (Table 2) to estimate $2c$ and $Y = 1.13$, the geometric factor for penny-shaped cracks. The assumption of penny-shaped cracks is based on characterization of BSE images, which show elongate pores are locally present in the optimum orientation for failure (e.g., Figure 4). Results are given in Table 4. These are only approximations of $K_I$, particularly for the St. Peter Sandstone samples, which do not show evidence of micro-cracks. However, they are comparable to reported values of $K_{IC}$ for sandstones, which range from 0.69 MPa m$^{1/2}$ to 2.40 MPa m$^{1/2}$ with lower porosity sandstones exhibiting higher $K_{IC}$ [Atkinson and Meredith, 1987]. Larger micro-cracks/pores than observed in thin section may exist within a given core, which would result in a larger $K_I$. These results suggest that samples were at or close to $K_{IC}$ early in the experiments.

Table 4. Experimentally Observed Time to Failure, $t$, and Results From Poroelastic Models

<table>
<thead>
<tr>
<th></th>
<th>$P_f$ (MPa)</th>
<th>$\sigma_3'$ (MPa)</th>
<th>$\sigma_{3p}'$ (MPa)</th>
<th>$\sigma_{3s}'$ (MPa)</th>
<th>$K_I$ (MPa m$^{1/2}$)</th>
<th>$\sigma_{3s}'/T$</th>
</tr>
</thead>
<tbody>
<tr>
<td>3_2</td>
<td>31</td>
<td>40</td>
<td>42</td>
<td>47</td>
<td>51</td>
<td>1.4</td>
</tr>
<tr>
<td>12_1</td>
<td>34</td>
<td>31</td>
<td>32</td>
<td>48</td>
<td>55</td>
<td>1.8</td>
</tr>
<tr>
<td>5_5 ⊥</td>
<td>1.5 &lt; $t$ &lt; 1.75</td>
<td>12</td>
<td>8</td>
<td>8</td>
<td>11</td>
<td>0.7</td>
</tr>
<tr>
<td>5_5</td>
<td></td>
<td></td>
<td>12</td>
<td>18</td>
<td>2.2</td>
<td>8</td>
</tr>
<tr>
<td>Abo ⊥</td>
<td>256</td>
<td>20</td>
<td>23</td>
<td>43</td>
<td>45</td>
<td>2.9</td>
</tr>
</tbody>
</table>

$P_f$ is the modeled pore pressure near the failure plane at the time of failure. $\sigma_3'$ is the effective stress in the vicinity of the failure plane at the time of failure determined by poroelastic models. $\sigma_{3p}'$ is the maximum effective tensile stress experienced near the failure plane calculated by models of the experiment and $\sigma_{3s}'$ is the maximum effective tensile stress experienced anywhere within the sample as calculated by the models. $\sigma_{3p}'$ and $\sigma_{3s}'$ occur immediately after the modeled axial load is reduced. The largest pores measured from BSE images of the samples (Table 2), were used to calculate the stress intensity, $K_I$, immediately following reduction of the axial load. The greatest effective tensile stress experienced by a sample exceeded its tensile strength by the factor $\sigma_{3s}'/T$. 3_2 is the isotropic sample and 12_1 contains concretions.

Figure 10. (a–e) Fracture locations and orientations relative to $p$ (top) and $\sigma_3'$ (bottom) at the time of failure. In the latter diagrams, blue indicates the most tensile stress. Axial strain distributions mimic the axial stress distributions and, therefore, are not shown.
(the isotropic sample) where tensile stress was predicted by the models to be greatest early on and at the time of failure (Figure 10a). In sample 12_1 (Figure 10b), neither the throughgoing nor the non-throughgoing fractures developed at the sample center where poroelastic models predict the greatest tensile stress at failure. In Abo (Figure 10e), like 12_1, the fracture did not form where the pore fluid pressure and tensile stress were predicted to be the greatest by the poroelastic models. In both 5_5 ⊥ and 5_5 || (Figures 10c and 10d) deformation bands formed near the core ends where the tensile stress was small.

Poroelastic models do not address deviations from linear elastic deformation due to non-linear elastic moduli and/or grain-scale deformation that alters the mechanical and hydrologic properties of the bulk rock. We therefore compared axial and radial strain leading to failure in the experimental cores with strain predicted by the models to evaluate how well the linear-elastic plane-strain models predict deformation under experimental conditions. The experimental axial elongation is plotted along with the total axial elongation predicted by the poroelastic models as a function of time in Figure 11. Poroelastic models predict rapid axial elongation followed by gradual shortening. However, the cores initially extended more slowly than the models predicted and continued to extend throughout the experiments. At the time of failure, most samples had experienced less total axial elongation than predicted by the models. Abo was the only sample to fracture after more axial elongation than predicted by the model. This sample followed the same pattern of gradually increasing axial elongation as the others, but took the longest to fail, which allowed it to exceed the predicted axial elongation (Figure 11).

Poroelastic models predict that core radii will rapidly shorten as sample axes extend (Figure 12). The proportionality between radial and axial strain is constant for each sample and is determined by the Poisson’s ratio, \( \nu \). It is therefore not surprising that the poroelastic models overestimate initial axial elongation just as they overestimated initial radial shortening. We would expect the discrepancy between modeled and experimental radial strain would be related to the discrepancy between the modeled and experimental axial strain by a factor of \( \nu \). However, the discrepancies in radial strain are much greater than those for axial strain, suggesting that deformation leading to failure during the undrained tests deviated significantly from linear elastic behavior.

8. Interpretations

Samples 3_2 (isotropic sample), 12_1 (sample with concretions), and Abo are inferred to have failed through production of opening mode fractures based on the orientations of the fracture surfaces and lack of evidence that indicates shearing. All produced fractures perpendicular to core axes, the orientation expected for a mode I fracture with the imposed stresses. The throughgoing fractures propagating through the cores only after significant sample dilation, at negligible pore pressure. However, none of the cores failed during the drained experiments, indicating that the differential stress imposed in the absence of elevated pore pressure was not sufficient for either fracture or deformation band formation. Therefore, we conclude that the structures formed during the NHF experiments are fluid driven. In addition, dilatancy was significantly greater during

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**Figure 11.** (a–e) Axial elongation as a function of time for experiments (solid lines) and for the poroelastic models (dashed lines). At time \( t = 0 \) s, pore fluid pressure and \( \sigma_a \) were reduced at the sample ends. The experimental samples initially extended less quickly than predicted by the models then continued to extend throughout the experiment. The models predicted that after initial extension the samples would begin to shorten, which was not observed experimentally. Differences between the experiments and models at \( t = 0 \) s reflect deviations from linear elastic behavior during the loading stage of the experiment.
the undrained tests than the drained tests and far exceeded the magnitude of dilation predicted by poroelastic models, suggesting that elevated pore pressures caused significant grain-scale damage that led to failure. [52] Samples 5_5⊥ and 5_5∥ also experienced significant dilation compared to their drained tests. However, the structure sets that formed are most likely hybrid deformation bands, meaning they show evidence of forming by a combination of extension and shear failure. In both cores, deformation bands formed at high angles, but not perpendicular, to the core axis and the maximum tensile stress. The precise relationship between principal stress orientations and hybrid fracture orientation is not well understood. Hybrid fractures in Carrara marble, however, form at lower angles to the maximum principal stress than pure shear fractures, which typically form at 20° – 40° to the maximum principal stress [Engelder, 1999; Ramsey and Chester, 2004]. Extrapolating these observations to our samples, we would expect hybrid structures to form more than 70° to the core axis, consistent with our data. However, it is not clear if this is the case because spatial variations in dilatancy, which would modify the pore pressure distribution, are not known. That these samples failed with hybrid deformation bands and the others did not is probably the result of differences in both their strength and porosity (Table 3).

8.1. Controls on Fracture Initiation and Propagation

[53] Cores experienced the greatest tensile stresses immediately after \(\sigma_a\) was reduced; however, they extended throughout the experiments, ultimately experiencing axial strains between \(2 \times 10^{-3}\) and \(2 \times 10^{-2}\), and required between \(\sim 20\) s and 250 s to form throughgoing structures. Thus, they ultimately failed at lower tensile stresses than they initially experienced. These results indicate that the effective stress and critical stress intensity criteria do not adequately predict the conditions leading to NHF formation, specifically under the high differential stresses and rapid deformation experienced in these experiments.

[54] Core dilation was between 4 and 12 times greater in undrained than in drained tests, indicating that elevated pore pressures significantly enhanced volume strain. We assume that the poroelastic models accurately predict pore pressure and effective stresses early in the experiments; however, we suggest they become less accurate as the modeled experiment progresses because they do not incorporate inelastic and non-linear deformation. As a result, the magnitude of the experimental volume strain is such that pore pressure must have been negligible prior to and at failure even though the poroelastic models predict that pore pressures and effective tensile stresses should be high even at failure (Figures 5 and 10). From these observations, we conclude

![Figure 12. (a–c) Radial strain as a function of time for experiments (solid lines) and for the poroelastic models (dashed lines). At time t = 0 s, \(\sigma_a\) and the pore pressure at the sample ends were reduced. The models predicted both much greater radial shortening at the start of the experiment than was observed experimentally and a greater increase in radial shortening as the experiment progressed. Differences between the experiments and models at t = 0 s reflect deviations from linear elastic behavior during the loading stage of the experiment.](image-url)
that although samples initially experienced effective tensile stresses high enough to meet the effective stress criterion, they first responded to the stress conditions by dilation rather than immediate failure. With respect to modeled strain, experimental dilation is characterized by inhibited radial shortening rather than enhanced axial elongation (Figure 11).

[55] In some cases dilation reduced the pore pressure of the system and the core eventually failed; however, in others the cores dilated without failure. Cores 3_2 and 12_1 dilated, but did not produce throughgoing fractures when exposed to effective tensile stresses of ~38 MPa, much greater than their ~1–2 MPa tensile strengths. They both produced throughgoing fractures only when the effective stresses were increased 10 MPa. Similar behavior was observed in 5_5. During both the lower and the higher pore pressure experiments, the samples dilated until the pore fluid pressure was low or negligible. Because equation (4) does not consider fluid flow out of the sample, it is not clear from Figures 5a, 5b, and 5d whether or not the tests that did not produce fractures reached zero pore fluid pressure, as these are lower bounds. However, we believe that it is likely that they did, since flow out of the sample was significant. The magnitude of mean stress, differential stress, and dilation thus were the primary differences between the cores that did not fail and those that did (Figure 5).

[56] Dilation was most likely accomplished by stretching pores and creating and propping open microcracks sub-parallel to the core axis as well as by closing pores and microcracks perpendicular to the core axis. Evidence for this is primarily provided by axial and radial strain patterns. Although axial extension of the cores was slower than predicted by the poroelastic models, the radial strain was negligible and inconsistent with a near-constant Poisson’s ratio, suggesting that radial shortening was inhibited as the cores extended. Since the undrained tests led to failure, whereas the drained tests did not, we interpret that fluid driven dilatancy and associated grain-scale damage, rather than differential stresses alone, ultimately caused fracture formation. The fact that pore pressures in the NHF tests were negligible at failure suggests that fluid-driven strain accumulation and not effective stress at the time of final fracture propagation through the cores was the primary control on failure. Although we were not able to monitor strain at crack tips during the localization process, our results support the assertion of Bruno and Nakagawa [1991] that elevated pore pressures can contribute to the strain energy of a system.

[57] The stress intensity factors estimated from stresses determined by the poroelastic models immediately following the reduction of the axial load and the maximum measured pore/microcrack lengths indicate that samples were at or close to their critical stress intensities (Table 4). Although sample dilation eventually reduced pore pressures below the values predicted by the poroelastic models, measured strains suggest that immediately after the axial load was reduced, before significant dilation, pore pressures and effective stresses were similar to those predicted by the models (Figure 11). However, since the critical stress intensities for these specific samples are estimates and stress distributions for porous rocks can be complicated, it is unclear if the cores were at or just near their critical stress intensities. The pore pressures within the cores decreased due to both fluid flow and pore volume expansion, so either the samples ultimately failed at lower stress intensities or they experienced deformation that increased the largest flaw size and therefore the stress intensity over time, thereby leading to failure.

8.2. Controls on Fracture Localization

[58] Although the rock behavior leading to fracturing observed in these experiments has not been previously described, the locations of experimentally produced fractures are consistent with theoretical and field evidence that fractures nucleate on mechanical heterogeneities. The most homogeneous and isotropic sample (3_2) fractured where pore pressure and the effective tensile stress were highest during the experiment; however, none of the other samples did. In all other samples, structures appear to have nucleated along mechanical and hydrologic heterogeneities. We infer that the fracture in Abo \(\perp\) nucleated on a plane of mechanical weakness, which is supported by observations of higher permeability along this plane during sample saturation. An iron oxide concretion at the edge of the throughgoing fracture surface of 12_1 is the most likely candidate for a point of nucleation in this sample.

[59] In addition to influencing the location of fracture initiation, results suggest that the nature of the heterogeneity influences the character of the structures produced. Abo \(\perp\) failed along bedding laminae, a recognized mechanical weakness, whereas Abo \(\parallel\), for which the same laminae were parallel to the tensile stress, did not fracture under the same stress and pore pressure conditions. The 5_5 samples demonstrated a similar anisotropy in tensile strength. The deformation band sets that developed in the two 5 5 cores were similar; however, 5_5 \(\parallel\) required greater tensile stress to fail. Thus, bedding created a mechanical anisotropy in these and the Abo siltstone samples by increasing the bulk strength of the rock perpendicular to bedding planes. Bedding laminae in Abo \(\perp\) produced spatially variable mechanical properties that reflect differences in strength of individual laminae. The 5_5 samples, which are much weaker than the Abo siltstone samples, did not, however, show evidence of deformation band localization along beds. Deformation bands in both 5_5 samples were produced in the same locations and orientations relative to the core, but not relative to bedding, and therefore appear to have been controlled by stress and pore pressure conditions.

[60] Only 12_1 had non-planar heterogeneity, and the resulting mesoscopic damage was more distributed in 12_1 than in the other samples. Specifically, it was the only sample to produce mesoscopic non-throughgoing fractures, which presumably localized on iron oxide concretions. Thus, by providing multiple nucleation sites for fractures, non-planar distributed heterogeneities may increase the damage a rock sustains in the presence of elevated pore pressures. This damage could increase bulk permeability and decrease rock strength.

[61] Sample 3_2 is the best indicator of hydrologic and dilatant controls on failure, as it is both as mechanically and hydrologically homogeneous and isotropic as is likely to be found in nature. The strain response of this sample to experimental conditions was similar to that of 12_1, and it failed at approximately the same time (i.e Figure 5). Failure occurred at its center where models predict the pore fluid pressure would have been greatest. Although poroelastic models do not incorporate the effects of locally reduced pore
fluid pressure due to dilation, we suspect that pore fluid pressure was greatest at the sample center in the initial stages of the experiment (prior to significant dilation) as predicted by the models. This probably led to greater dilation and damage near the sample center.

8.3. Relevance to Geological Processes

Our experimental results can be used to better understand the mechanics of natural hydraulic fracture formation. Although the experiments were designed to meet the NHF failure criterion and not to simulate the conditions of a specific geologic environment, our results do provide new insight into processes of tensile crack initiation and propagation. Experiments necessitated an abrupt increase in differential stress and therefore high strain rates. Fault zones, in particular, experience spatially and temporally varying strain rate and differential stresses, and are therefore the most likely location of natural examples of the process we explored here.

Fault zone stresses vary over the course of the seismic cycle. Where pore fluids cannot communicate with the surface fault zones experience pre-seismic increase in both differential stress and pore fluid pressure at very low strain rates [Sibson, 1992]. When seismic rupture occurs it abruptly reduces differential stress in the primary slip zone. Dynamic stress changes during passage of the rupture front, however, have been proposed to produce tensile cracks in the fault damage zone [Dalguer et al., 2003]. Such dynamic changes, involving a rapid increase in magnitude of tensile stress, if coupled with essentially instantaneous drainage of elevated pore fluid pressure within the fault zone, might be the best natural example of the experiments reported here.

The strain rates and pore pressures in these experiments were dictated by the requirement that pore pressure be greater than the least compressive stress during the initial stages of the experiments. The period of time over which this condition could be maintained was limited by the sample length and diffusivity. In nature, the length scale of the system will be much greater and, therefore, the pore pressure may not decline as quickly. When pore pressures are sustained over longer timescales, the damage imparted by the pore pressure may be greater for lower pressures, differential stresses, and strain rates, but determining the relationships between these parameters requires significantly more data.

9. Conclusions

We have documented patterns of failure in extension of clastic sedimentary rocks with a range of mechanical anisotropy and heterogeneity, with hydraulic conductivity fixed by pore fluids of variable density. In contrast to rupture tests, the location of fracture nucleation and propagation was not dictated by our experimental set-up, but instead was determined by the physical properties of the samples, and is therefore more relevant to hydraulic fractures produced in nature. The fractures formed nucleated on mechanical heterogeneities and, in the absence of heterogeneity, nucleated where pore pressure was initially greatest, consistent with both theoretical and field-based models of NHF formation. These observations suggest that we successfully reproduced the fundamental natural processes that control NHF initiation.

Fractures observed in the field are commonly interpreted to be NHF that propagated when the pore fluid pressure overcame the tensile strength of the rock, or using a micromechanical model, the critical stress intensity factor. However, our experiments demonstrate that the mechanics of fracture formation is more complicated than these stress criteria suggest. In the experiments reported here, pore pressure was quickly reduced by sample dilation. Therefore, under rapid deformation conditions, the initial pore pressure required to fracture rock was much greater than previously realized. The porous rocks in these experiments responded to experimental conditions by dilating, which reduced pore fluid pressure to negligible values. Throughgoing fractures are interpreted to have propagated through cores when dilation created sufficient damage.

Mesoscopic mechanical heterogeneity in the form of either planar discontinuities or distributed inclusions was the first-order control on fracture initiation in these experiments. However, mesoscopic heterogeneity was not required to nucleate fractures and, in its absence, fracture initiation was localized where pore fluid pressure was both greatest and elevated for the longest period during the experiment. The specific character of the heterogeneity affects both the location of throughgoing NHF and the distribution of damage. Whereas planar heterogeneities localize fracture nucleation along bedding laminae, distributed mechanical heterogeneity results in distributed damage in the form of non-through-going fractures. This distributed damage has the potential to impact both the hydrologic and mechanical properties of the bulk rock by increasing both permeability and compliance.

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References


