Role of permeability and storage in the initiation and propagation of natural hydraulic fractures

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Joint sets within sedimentary basins are commonly interpreted to have formed by tensile failure in conditions where pore fluid pressure was elevated. Such tensile fractures are inferred to be a part of the process that relieves high fluid pressure by locally increasing rock permeability. In spite of the importance of this feedback mechanism, the detailed mechanics of hydraulic fracture genesis remain poorly understood. We describe the results of both experimental and numerical studies of hydraulic fracture genesis on the basis of an experimental protocol that combines rock extension with elevated pore fluid pressure such that the hydraulic fracture criterion is met in the sample interior. This is achieved by simultaneously dropping both minimum stress and external pore fluid pressure, inducing a large fluid pressure drop between the sample interior and its ends. Poroelastic modeling suggests that the pore fluid pressure is highest close to, but not at, the sample ends and is locally maintained at levels that meet the hydraulic fracture criterion for up to 50 s. Application of this experimental protocol to an impure sample of sandstone resulted in the generation of several hydraulic fractures subparallel to the maximum principal stress. Fracturing did not occur in a drained test on the same sample, demonstrating that the elevated pore fluid pressure was critical to fracture formation. To better understand the experimental results, we explore the role of rock permeability and storage on fracture processes using a numerical model that directly couples a lattice-Boltzmann model for fluid mechanics with a discrete element model for solid mechanics. Like the experiment, the numerical simulations produced opening mode fractures when operated to replicate the conditions of the experiment. Fractures preferentially occur in portions of the model inferred to be mechanically weak. Local fluid pressure gradients strongly influence the state of stress in the solids and thereby fracture growth. Increasing the model permeability increases fracture propagation rate, decreases sample deformation, and increases fracture spacing. Sample deformation increases, and fracture spacing decreases, with increasing overpressure. It appears that bulk forcing of the solid via fluid seepage forces is important in fracture genesis, explaining the key roles of permeability and diffusivity in the hydrofracture process.


1. Introduction

Pore fluid pressure within the Earth’s shallow crust can exceed that predicted by the hydrostatic gradient [e.g., Neuzil, 1995], locally reaching magnitudes high enough to fracture rock. The generation of natural hydraulic fractures in response to elevated pore fluid pressure has been recognized and discussed for over 40 years [Secor, 1965], yet the conditions under which such fractures form remain sufficiently poorly understood that we lack the ability to model either their location or their density in, for example, an evolving sedimentary basin. Thus, although structural geologists commonly interpret extension fractures found in sedimentary basins as natural hydraulic fractures [Engelder and Lacazette, 1990; Cruikshank et al., 1991; Capuano, 1993; Foxford et al., 2000; Bahat et al., 2003] and hydrologists call on natural hydraulic fractures to mitigate high pore fluid pressure in basins through increased permeability [Bradley and Powley, 1994; Roberts and Nunn, 1995; L’Heureux and Fowler, 2000; McPherson and Bredehoeft, 2001], an explicit connection between hydraulic and mechanical processes in fracture formation is lacking. The goal of the research reported here is to enhance understanding of the intrinsic and extrinsic conditions under which natural hydraulic fractures form, which will allow better estimates of their extent and importance in the subsurface.

Geologic controls on fracture spacing in sedimentary rock systems are not quantitatively well understood. Nelson
[1985] provided a summary of accepted controls that includes (1) rock composition, (2) rock “texture” (including grain size and porosity), (3) structural position within a basin, and (4) mechanical stratigraphy. The most commonly cited, and arguably best documented, geologic control on fracture spacing in sedimentary rocks is mechanical layer thickness. Mechanical layers are discrete mechanical entities, most commonly sedimentary beds sandwiched between physically unlike beds or weak bedding planes. Fracture spacing is observed to decrease with mechanical layer thickness [Price, 1966; Ladeira and Price, 1981; Huang and Angelier, 1989; Narr and Suppe, 1991; Gross, 1993; Becker and Gross, 1996; Gross et al., 1997; Bai and Pollard, 2000a]. Documented ratios of fracture spacing to bed thickness range from 12 to 0.01, with purely mechanical models explaining ratios of 0.8 to 12 as the result of extensional boundary loading and subsequent fracture saturation [Bai and Pollard, 2000a]. Closely spaced fractures, with ratios of <0.8, are inferred to be the result of elevated pore fluid pressure, but this influence has not been investigated using models that explicitly consider fluid overpressures.

[s] Renshaw and Harvey [1994] addressed the quasi-static growth rates of natural hydraulic fractures by simulating fractures in poroelastic media using a displacement discontinuity boundary element method. They noted that poroelastic effects may limit fracture growth rate, but growth may still accelerate. In their model, the growth rate of an isolated natural hydraulic fracture is typically a function of the characteristic time of a given rock. Characteristic time is dictated by hydraulic conductivity, which determines how quickly fluid is transmitted to the fracture. This is true because, unlike induced hydraulic fractures, the fluid pressure within natural hydraulic fractures is less than the ambient fluid pressure after propagation begins, as fluid pressure in the crack drops once the fracture opens and allows the pore fluid to expand.

[s] The difference between processes of natural hydraulic fracturing and induced hydraulic fracturing for enhanced geothermal and oil extraction is significant. The propagation of induced fractures explicitly depends on the stress distribution only around the fracture tip. Thus, many papers [e.g., Rice and Cleary, 1976; Advani et al., 1987] address pore fluid pressure at the crack tip. Some theoretical arguments underpredict the amount of pressure it takes to propagate an induced hydraulic fracture because a small region of reduced pressure at the tip of a propagating fracture can significantly reduce the stress intensity and the fracture therefore requires renewed abnormally high fluid pressure to propagate [Advani et al., 1997]. This lower-pressure region at the fracture tip, the fluid lag region, is usually attributed to excessive fluid leak off into the formation. This result suggests the converse for natural hydraulic fractures, namely that fluid flow from the formation into the fracture may be a rate-limiting process in natural hydraulic fracture propagation, just as Engelder and Lacazette [1990] have discussed and Renshaw and Harvey [1994] demonstrated.

[s] Petroleum engineering studies suggest that hydraulic fracturing may be affected by creating local regions of low effective stress via fluid injection. Bruno and Nakagawa [1991] demonstrated that induced hydraulic fractures will propagate toward regions of higher local pore pressure, or lower effective stress. The stress concentrations at crack tips are locally influenced by matrix pore pressure magnitude. In their study, higher injection pressures introduced greater deviation in fracture propagation direction. Berchenko and Detournay [1997] used a numerical model to demonstrate that matrix pore pressures can alter the propagation path of an induced hydraulic fracture.

[s] The research presented in this paper is a logical extension of the above-described work. We have integrated experimental and numerical studies to explore the relative controls of mechanical and hydrologic rock properties on hydraulic fracture formation. The different avenues of investigation, experimental deformation, poroelastic modeling, and discrete numerical modeling, are linked by a common experimental protocol that produces a pore fluid pressure gradient within a test sample or model space. Discrete numerical models show that fracture spacing decreases with permeability. This work suggests that rock permeability plays a key role in the formation of bed-confined hydraulic fractures in sedimentary rocks. This result may in part reflect the fact that low diffusivity allows high pore fluid pressure to be maintained for a longer period of time in a test sample or model space. Interestingly, both model and experimental fractures initiate not just where pore fluid pressure is sufficiently high to meet the fracture criterion, but at locations where it is at or near maximum in a given sample. The exact locations of fractures also are influenced by mechanical anisotropy and material heterogeneity. Discrete numerical models also suggest that seepage forces are important in overcoming confining forces to produce fluid pressure-driven extension fractures. Fracture spacing is therefore proposed to scale with both degree of fluid overpressure and permeability.

2. Natural Hydraulic Fractures

[s] Secor [1965] hypothesized that if the fluid pressure was greater than the least minimum stress \( \sigma_3 \), the effective minimum stress \( \sigma_3^* \) became tensile:

\[
\sigma_3^* = \sigma_3 - P_f. \tag{1}
\]

If this (tensile) effective minimum stress is greater than the tensile strength of the material, then new void space is created as tensile fractures propagate. This relationship, known as the hydraulic fracture criterion, represents a force balance between the pore fluid and the solid and is widely used throughout geology and hydrogeology. If it is correct, then either a decrease in \( \sigma_3 \) or an increase in \( P_f \) is needed to produce an opening-mode fracture in a rock.

[s] Another way to examine the force balance in equation (1) is to consider the resultant forces per unit volume acting on the solid portion of the porous medium [Wang, 2000]. Following Terzaghi [1943], three forces act on the solid particles through which flow occurs while ignoring the effects of lateral confinement forces. The first is produced by the weight of the overlying solids \( F_1 \) acting downward:

\[
F_1 = -\gamma_s (1 - n) \nabla z, \tag{2}
\]
where \( \gamma_f \) is the specific weight of the dry solid particles, \( n \) is the porosity, and \( \nabla z \) is the gradient in elevation (parallel to the gravitational field). The second is the buoyancy force \( F_2 \), defined as
\[
F_2 = -(1 - n) \nabla P_f,
\]
in terms of force per unit volume and \( \nabla P_f \) is the gradient in pore fluid pressure. The final force is the drag (or seepage) force \( F_3 \) which, per unit volume of porous medium, is equal to
\[
F_3 = -n \rho_f \nabla \psi,
\]
where \( \rho_f \) is the fluid’s density, and \( \psi \) is the fluid potential defined as
\[
\psi = z + \frac{P_f}{\gamma_f},
\]
where \( z \) is elevation and \( \gamma_f \) is the specific weight of fluid. The fluid potential is a measure of the available energy per unit weight of fluid and is dissipated as viscous friction at fluid-solid interfaces. As this energy is dissipated, a force is exerted on the solid matrix in the direction of fluid flow. The potential gradient can be attributed to the porous medium through the permeability or external driving forces depending on scale. Terzaghi [1943] neglected externally applied forces, but these can easily be incorporated through force summation over a volume of porous material. Note that if \( \nabla P = 0 \) (i.e., no flow) then \( F_3 = F_2 = 0 \). This suggests that in absence of a fluid pressure gradient no net force from the fluid (ignoring gravity) will be imparted to the solid framework.

[10] Mourgues and Cobbold [2003] similarly argue that the fluid pressure gradient exerts a tangible and important seepage force, and show convincingly in experiments on unconsolidated sediments that this total force is independent of fluid permeability. They suggest that fluid permeability influences the nature of stress (fluid pressure or shear stress) and the way it is distributed on both external and internal surfaces, but not the total stress on the rock or sediment. Thus permeability may be a control on localization of deformation in heterogeneous systems. Recently a paper by Cobbold and Rodrigues [2007] suggests that seepage forces are extremely important factors in the formation of horizontal hydraulic fractures.

[11] Significantly, the force relationships expressed above are derived per unit volume of porous medium, meaning that forces applied to the solid framework scale with the volume of interest. This implies that the forces per unit volume induced on the solid framework by a given pressure gradient are identical from the centimeter to the kilometer scale. Below we present an experimental technique that is capable of exploring the importance of the forces induced by such a fluid pressure gradient on the mechanical behavior of rock.

[12] The conditions of natural hydraulic fracturing are not simple to impose in a laboratory test. For example, dropping \( \sigma_3 \) while maintaining a higher pore fluid pressure is impossible in a standard cylindrical triaxial pressure cell. This is because these setups typically apply \( \sigma_3 \) with a loading piston axially attached to the end caps of the sample. It is inevitable that when \( \sigma_3 \) is dropped lower than \( P_f \), fluid will leak between the end caps and the rock, creating an “extension fracture” between them. An alternative would be to apply a minimum principal stress through a confining pressure surrounding the long axis of the specimen, but when \( \sigma_3 \) is dropped lower than \( P_f \), fluid will break the seal between the sample and the confining fluid. Without designing a new testing apparatus, dropping \( \sigma_3 \) and holding \( P_f \) constant is not possible. Increasing \( P_f \) while holding \( \sigma_3 \) constant suffers from the same set of problems.

[13] A new approach to generating hydraulic fractures was designed to overcome the limitations of laboratory boundary conditions. The test, shown in Figure 1, is a combination of the end-member cases discussed above. Initial conditions on a right cylindrical sample of rock saturated with a fluid pressure \( P_{fo} \) are typical of conventional triaxial tests, with the exception that \( \sigma_{3o} \) is axial and \( \sigma_1 \) laterally confines the sample. At the start of the test, both \( P_{fo} \) and \( \sigma_{3o} \) are instantaneously lowered to new values of \( P_f \) and \( \sigma_3 \), respectively. These conditions will not cause leaking at the boundaries or through the sample jackets since (1) \( P_f \) is lower than \( \sigma_3 \) at the boundary and (2) \( P_f \) is lower than \( \sigma_1 \) everywhere. The drop in both stress and fluid pressure at the boundary is intended to create conditions conducive to the generation of extension fractures in the sample: fluid pressure in the sample interior that is greater than the local axial stress which, assuming quasi-equilibrium with new boundary conditions, should be \( \sigma_3 \). As a result, extension fractures are expected to form perpendicular to \( \sigma_3 \) in the volume where \( P_f \) is high and \( \sigma_3 \) is low. Since the force induced by the fluid pressure gradient on the solid skeleton scale with volume, forces induced by a large fluid pressure gradient on a small sample are geologically realistic; the same pressure gradient over a large area would reveal a much larger force. Additionally, in natural hydraulic fractures, unlike induced hydraulic fractures, ambient fluid pressure is higher surrounding the fracture than inside the fracture itself, at least until fluid drains into the fracture and has time to equilibrate. This condition imposes several constraints on how resulting fractures initiate and interact during propagation.

[14] Differing time constants governing fluid flow and mechanical equilibrium are responsible for creating a condition of high \( P_f \) and low \( \sigma_3 \) at the sample center. Specifically, the fluid diffusion time constant is much lower (i.e., fluid diffusion is slower) than the mechanical equilibrium time constant. In porous media, the time it takes for a given fluid pressure to come to equilibrium with a change in boundary conditions is a function of the rock’s hydraulic diffusivity (\( \kappa \)) defined here as
\[
\kappa = \frac{k \rho g}{\mu S_s},
\]
where \( S_s \) is the one-dimensional specific storage, \( k \) is the intrinsic permeability, \( \rho \) is the fluid density, \( g \) is gravitational acceleration, and \( \mu \) is fluid dynamic viscosity. Note that \( \kappa \) has units of \( \frac{\text{length}^2}{\text{time}} = \text{m}^2/\text{s} \). A time of \( \frac{L^2}{\kappa} \) is required for a pressure change to propagate a distance \( L \). The diffusion
time is a function of the rock’s hydrologic properties \((k, S_s)\) and it depends on the fluid properties \((\rho, \mu)\).

3. Experiment Proof of Concept

[15] Sandstone from the Permian Abo Formation of central NM was used to test the experimental technique described above. The moderately well sorted, very fine-grained channel sandstone we sampled for this work is dominated by quartz, but includes significant clay in addition to some feldspar and local white mica and chlorite. Grain shapes vary from subrounded to subangular. The rock is very well indurated, due largely to quartz overgrowths but in part also to patchy carbonate (both dolomite and calcite) cement. Millimeter-scale bedding laminae are distinguished by variations in mineralogy, particularly clay and dolomite content. The tensile strength of the sandstone was measured at 4 MPa using a standard Brazil test with a 0.0254 m diameter sample. Permeability normal to bedding is \(10^{-17}\) m\(^2\), calculated by a steady state flow test. This relatively low permeability and assumed storage (porosity is roughly 12%) make this sandstone an ideal rock for this test, as these factors result in a low hydraulic diffusivity. As noted previously, the time required for fluid pressure to come to equilibrium with a change in boundary conditions is directly related to hydraulic diffusivity. For this test, the longer it takes for fluid to come to equilibrium (i.e., the lower the diffusivity), the longer the fluid pressure will remain greater than required by the hydraulic fracture criterion. A plot of the time to reach pressure equilibration versus rock hydraulic diffusivity provides an estimate of the lower limit at which the change in stress must take place for this criterion to be met (Figure 1c). The reaction time to the change in boundary conditions increases with fluid viscosity. The solid line placed at 3 s in Figure 1c is the assumed reaction time of the laboratory equipment. For a fluid with the viscosity of water, any rock with a diffusivity greater than 0.003 m\(^2\)/s will have a fluid pressure reaction time greater than that of the machine, and the fluid pressure distribution we describe will not be produced. The time constant (shown on plot) for Abo sandstone is well above this value; for a pore fluid viscosity 50 times that of water (the fluid used in this experiment), the time constant is roughly 200 s.

3.1. Poroelastic Modeling of Experimental Protocol

[16] Two-dimensional, axisymmetric, poroelastic numerical models executed with the BIOT2 code were used to better understand the poroelastic response of the Abo sandstone to changing laboratory conditions. The Abo model is compared to a model constructed with equivalent mechanical properties but higher intrinsic permeability; both are contrasted with a case in which the system is drained throughout the experiment, and therefore fluid pressure remains close to atmospheric pressure. BIOT2 is a finite element model developed by Paul Hsieh of the USGS that solves the governing equations of fluid flow and deformation in elastic media (using the equations of linear poroelasticity). Boundary conditions for the mechanical and hydrologic components of the axisymmetric model are presented in Figure 2. Model input parameters for the
coupled Abo sandstone model consist of both mechanical and hydrologic properties listed in Table 1. For the purposes of these models, the sandstone is assumed to be mechanically and hydrologically homogeneous. We know this is not the case in the natural sample; however, the observed millimeter-scale variations in physical properties are too small to quantify and unreasonable to incorporate into a numerical model built on continuum assumptions. Simulations began with isotropic loading to the conditions to be imposed in the laboratory test followed by the reduction in \( P_f \) dictated by the test design.

[17] Plots of axial displacement of the sample as a function of time show that the low-permeability material undergoes significantly more elastic axial extension (negative displacement) than high-permeability material for the same, undrained test protocol (Figure 3a). A drained (i.e., no change in fluid pressure at boundaries), low-permeability system experiences about three times less axial extension than the same low-permeability material with elevated fluid pressure. In addition, the lower-permeability, undrained test protocol (Figure 3a) experiences about three times less axial extension rate for a significantly longer time than the higher-permeability model.

This phenomenon suggests that the \( P_f \) plateau reached in the first \( \sim 30 \) s of the test (gray line in Figure 3b) reflects drainage of fluid inward toward the sample center. This flow causes the fluid pressure within the sample to be elevated for a longer period of time than would be possible if dilation were homogeneous throughout the model domain, and causes the break in slope in both sample displacement and pore fluid pressure observed in Figure 3. The lower the hydraulic diffusivity of the sample, the longer a relatively high \( P_f \) can be maintained, allowing sample extension to continue, even in a purely poroelastic model which does not account for inelastic deformation mechanisms such as fracturing.

### 3.2. Experimental Setup

[19] As described previously, the experimental design requires the application of a confining maximum stress.
Figure 3. (a) Axial displacement and (b) pore fluid pressure evolution of three poroelastic models. Solid and dotted lines show the time evolution of axial displacement and pore fluid pressure following a 15 MPa $P_f$ drop with the properties of Abo sandstone ($k = 1 \times 10^{-16}$ m$^2$) and a model with identical mechanical properties but a higher permeability ($k = 1 \times 10^{-13}$ m$^2$), respectively. The dashed lines reflect a drained case where an Abo sandstone sample is saturated but no fluid pressure gradient is imposed. Time indicated on the plots is relative to drop in $P_f$ and $\sigma_3$. Only the low-permeability (and low-diffusivity), undrained model experiences significant sample extension for an extended period of time; extension is coincident with the period of elevated pore fluid pressure.

Figure 4. (a) Contour plot of pore fluid pressure in the Abo sandstone poroelastic model 1 s after start of the saturated extension test. Pore fluid pressure peaks are located near model “sample” boundaries because of sample volumetric dilation (pore space increase) and the Skempton effect. (b) A time comparison of $P_f$ for a node near the sample boundary (dashed line) and a node at the sample center (solid line) within the Abo sandstone poroelastic model. The former nodes have the highest $P_f$ in the model, although much of the sample has sufficiently high pore fluid pressure to meet the hydraulic fracture criterion throughout the test. The node located in the center of the model indicates that pore fluid pressure remains steady at just over 12 MPa for $\sim$20 s while receiving fluid flow from the boundary regions.
and an axial minimum stress (Figure 1). This approach necessitates a special equipment design that allows such a differential stress to be maintained at the end caps (see auxiliary material). The end caps allow significant axial extension while preventing confining fluid from infiltrating the sample. Fluid ports bored through the piston and into each end cap allow the external maintenance of a pore fluid pressure. A distributor plate is placed between each end cap and the sample to allow even distribution of the pore fluid across the face of the sample. The specimen was jacketed with a polyolefin heat shrink tubing and coated with a polyurethane epoxy to secure the sample end cap assembly. The purpose of the epoxy and tubing is to provide guidance for the sample to remain aligned with the end cap. The sample was also equipped for measuring axial and lateral displacements via linear variable differential transducers (LVDTs). Two LVDTs placed along the long axis of the sample measure axial displacements and a circumferential split ring enables an additional LVDT to measure circumferential displacements. Tests described in this paper were performed at the geomechanics lab at Sandia National Laboratories.

The Abo sandstone was cored with its millimeter-scale bedding laminae oriented about 80° from the long axis of the test sample. This geometry is such that if laminae are zones of weakness, extension fractures might form in the sample from purely mechanical unloading, in the absence of an elevated pore fluid pressure. To test this possibility, the sample with end cap assembly was placed in the pressure vessel and a 1 MN load frame. It was loaded isotropically to 20 MPa, after which the axial load was lowered to 18 MPa. A drained (i.e., no pore fluid pressure gradient) run then was executed, where the sample was first loaded isotropically to 20 MPa, after which the axial load was dropped abruptly to 0 MPa. This was done to determine if a differential stress of 20 MPa was large enough to induce a fracture in the absence of an elevated pore fluid pressure. It was not. Subsequent to this test, the sample was prepared for a pore fluid pressure gradient test. A silicone-based pore fluid (Dow 250 -5 × 10 m/s viscosity) was pumped into the bottom of the sample, maintaining a 10 MPa pressure drop across the sample, until breakthrough was observed at the top. Pump directions were then reversed and fluid was pumped until the sample reached 15 MPa. Valves were attached to the top and bottom inlets to the pressure vessel, to allow for the simultaneous lowering of Pf to atmospheric pressure at both ends of the rock sample at test initiation.

3.3. Test Results and Interpretation

The test apparatus would optimally allow both fluid pressure and axial stress to be lowered simultaneously so that pore pressure is larger than the minimum stress in portions of the sample. Because this is not possible with the available equipment, we used the slow fluid pressure equilibrium time of the sample to our advantage. Pf was dropped before axial stress to ensure that the sample jacket remained intact. The time between drops was ~1 s. Figure 5 shows the experimentally imposed changes in fluid pressure, confining pressure, axial stress, and resulting sample extension over the course of the test.

Before the test was initiated, all parameters were maintained at a constant level. To begin the test, pore fluid pressure was lowered by opening the valves of the pore fluid ports to the atmosphere. Axial stress was then reduced by moving the loading platen about 1 mm vertically upward causing the sample to extend but remain in contact with the sample end caps (Figure 5). This initial axial stress reduction resulted in most of the sample extension and, presumably, the formation of extension fractures at some scale, though none were throughgoing. After 10 s the loading platen was moved an additional 0.1 mm because extension was causing the sample to push against the end caps. An audible “pop” that coincided with this additional unloading event suggests instantaneous opening of the sole throughgoing extension fracture formed in this experiment. This could represent either final propagation of the fracture across the sample or an instantaneous increase in aperture. Approximately 15 s after this movement, a sharp jump in axial stress occurred with no corresponding platen displacement, which we interpret to mark the time at which this extension fracture was sheared because of the differential stress on the fluid-filled fracture (arrow in Figure 5; the figure is discussed further below). The draining of fluid coupled with sample strain resulted in an increase in axial stress even as the platen was unloading the sample. The rock subsequently was allowed to drain for 4 h before the sample assembly was removed from the pressure vessel.

A plot of axial stress versus axial strain highlights the above described features of the test (Figure 6). Four times more strain took place in the pore fluid pressure gradient test than in the drained test. No evidence of inelastic strain, or fracturing, was observed in the drained test. Thus, elevated fluid pressure allowed the sample to accumulate substantially more strain (elastic and inelastic) in the fluid pressure gradient test than in the drained test. The three increases in axial stress subsequent to test initiation corre-
The sample boundaries (Figure 7b). Instead, they were generated closer to the rock. None of the extension fractures formed in the throughgoing fracture, with the remaining rock consisting of albite feldspar, dolomite, clay, and calcite, in order of decreasing abundance. In contrast, the fracture surface is less than half that the sandstone is roughly 3/4 quartz adjacent to the laminae above and below the fracture. The maps show X-ray maps of the throughgoing fracture surface to maps of the laboratory sample, in particular, might be explained by extension fracture (3rd from sample bottom) observed in the latter test.

Comparison of model with experimental results suggests that the extension fractures formed in response to elevated pore fluid pressure. More specifically, although most of the sample met the hydraulic fracture criterion at some point during the experiment, fracturing preferentially took place in regions that experienced the highest fluid pressure, which exceeded that necessary to meet the criterion.

Since the sample failed near its ends, in regions where high $P_f$ likely did not last very long (see Figure 4b), the fractures most likely formed in a relatively short time interval following the start of the test. After the initial fractures formed, it is possible that they altered the pore fluid pressure history of the sample such that it followed a path different from the pressure distribution estimated from the simulation. An increase in permeability through fracturing near the sample ends would mean that the highest pore fluid pressure would be found in the center of the sample. The anomalous extension fracture (3rd from sample bottom) observed in the laboratory sample, in particular, might be explained by such a “wave” of elevated pore fluid pressure, with the fractures closest to the boundaries forming first and fractures closer to the center occurring later.

All of the fractures produced in the experiment are parallel to bedding laminae oriented 80° from the long axis of the specimen. Fractures therefore nucleated on these heterogeneities; the heterogeneities are not, however, sufficient to facilitate fracture formation in the absence of high pore fluid pressure, as demonstrated by the results of the drained experiment in which no fractures were produced. We deduce from these observations that optimally oriented physical heterogeneities assisted in the localization of failure, but were insufficient to cause fracturing in the absence of elevated pore fluid pressure. We do not, however, have enough data to evaluate the relative roles of mechanical versus hydrologic variations associated with these laminae (i.e., small-scale variations in mechanical properties versus diffusivity).

**Figure 6.** Axial stress versus axial strain for pore fluid pressure gradient and drained extension tests. In both cases, significant extension (negative strain) of the rock occurred as the sample was unloaded. However, the strain in the fluid pressure gradient test sample was four times greater than that in the drained test. Both drained and fluid pressure gradient tests accumulated most of their sample extension at low axial stresses. A 1 MPa increase in the axial stress marks the time at which a major extension fracture was sheared in the latter test.

spond to the increases depicted in the time plot of Figure 5. Most of the sample extension took place during the first few seconds of unloading (compare Figures 5 and 6).

Rock extension was accommodated by both elastic strain and the initiation and propagation of extension fractures. Most fractures did not traverse the sample, but the sample was split by one throughgoing extension fracture (Figure 7a). All of these fractures are oriented 80° from the long axis of the specimen, parallel to bedding laminae. To evaluate mineralogic controls on fracturing, we compared X-ray maps of the throughgoing fracture surface to maps of the laminae above and below the fracture. The maps show that the sandstone is roughly 3/4 quartz adjacent to the fracture, with the remaining rock consisting of albite feldspar, dolomite, clay, and calcite, in order of decreasing abundance. In contrast, the fracture surface is less than half quartz, with greater amounts of feldspar, dolomite, and clay than adjacent areas. Feldspar and clay are mechanically weaker than quartz at shallow crustal conditions, suggesting the throughgoing fracture followed a plane of weakness in the rock. None of the extension fractures formed in the center of the sample; instead, they were generated closer to the sample boundaries (Figure 7b).

**3.4. Discussion of Experimental Results**

Throughout the duration of the test, over most of the sample, $P_f$ was greater than both $\sigma_3$ and the tensile strength of the Abo sandstone (4 MPa, as mentioned earlier), indicating that the hydraulic fracture criterion was met everywhere except at the very ends of the sample. This implies that the sample had an equal potential to fail anywhere along its length in a direction perpendicular to $\sigma_3$, assuming a uniform tensile strength. The laboratory results, however, show that deformation was localized closer to the sample ends than to its midpoint, and therefore was not equally distributed along its length (Figure 7b).

These observations could be explained by zones of mechanical weakness (where tensile strength is small), but because the fracture criterion was met throughout the sample interior, this is not a satisfactory explanation of the laboratory observations. An alternative explanation is provided by the poroelastic modeling results.

Although our poroelastic models do not simulate the inelastic deformation (i.e., fracture formation) observed in the laboratory samples, they provide a more complete framework in which to interpret observations. For example, the identification of zones in the sample where the pore fluid pressure meets or exceeds the hydraulic fracture criterion yields insight into the localization of failure. The Abo sandstone models show the highest pore fluid pressures attained during the numerical experiment result from volumetric compaction, causing pore pressure to rise near the sample ends; in these areas, $P_f$ exceeds $P_{fo}$ (Figure 4a). These regions of higher-than-initial pore fluid pressure roughly coincide with the locations of fractures formed in the laboratory tests, near the sample ends, with the exception of the third fracture from the sample bottom (Figure 7b).
The throughgoing extension fracture is interpreted to have sheared after it formed. The fracture was presumably fluid filled, and therefore had virtually no shear strength. Thus, the small component of shear stress operating on the fracture surface was sufficient to cause slip, in spite of its $10^\circ$ angle to the maximum compressive stress. Previous study of compressive failure of anisotropic rock showed that the shear strength of rocks is much lower where mechanically weak foliation planes are available to accommodate the deformation [Donath, 1961]. We infer that the fluid-filled fracture was a similar plane of weakness, albeit for a different reason.

4. Discrete Element Models of Fracturing

As noted earlier, poroelastic modeling captures only the elastic response of granular porous media to changes in stress and pore fluid pressure, and therefore does not allow us to analyze processes occurring during fracture formation. Deformation experiments provide a record of the stress-strain history of combined elastic and inelastic deformation, but do not provide a mechanism for exploring the dynamic interaction between hydrologic and mechanical processes over the history of the experiment. To evaluate these dynamic interactions over time, we have used a direct simulation method developed by Cook et al. [2000] and Cook [2001]. This method couples a discrete element model (DEM) for solid mechanics with the lattice Boltzmann method (LB) for fluid mechanics. Since the LB approximates the Navier-Stokes equations of fluid flow, no empirical relations (i.e., Darcy’s law) are necessary to relate fluid flux to pressure drop at the pore scale. Also, the DEM does not require a priori assumptions about the relationship between macroscale properties such as fluid permeability and porosity. The hydrodynamics develop with the evolving solid matrix. Details of this method [Boutt et al., 2007] are briefly reviewed below and are available in the auxiliary materials associated with this paper.

4.1. Discrete Element Method

DEM techniques have been used successfully to approximate the behavior of noncohesive, granular systems under low-stress conditions [Cundall et al., 1982] and lithified sedimentary rocks [Bruno and Nelson, 1991; Potyondy et al., 1996; Hazzard et al., 2000; Boutt and McPherson, 2002]. In this study, we employed an existing
two-dimensional DEM application [Rege, 1996]. A DEM simulates the mechanical behavior of a porous medium by idealizing the system as a collection of separate particles that interact at their contact points. The method consists of (1) identifying elements in contact and (2) resolving the contact physics. The calculations performed in the DEM alternate between the application of Newton’s second law and a force displacement law (simple contact models) at the contacts between particles. The force displacement law relates components of force to the corresponding components of the relative displacements through a generalized contact constitutive model. The contact constitutive model has two parts, a stiffness model and a slip model. The motion equations are integrated explicitly with respect to time to obtain particle positions, then used in the force displacement calculations, and the calculation cycle starts over again. An advantage of this approach is that the DEM constitutive behaviors (stress and strain relations) are results rather than assumptions.

4.2. Lattice-Boltzmann and Coupled Model Theory

[32] Cook [2001] and Cook et al. [2004] coupled fluid flow with mechanical modeling through the integration of LB with the DEM framework described above. The fluid lattice is a background Eulerian framework over which the Lagrangian solid discrete elements interact. Discrete elements “float” over the fluid lattice, with the position of the elements are calculated as a percentage overlap on the fluid lattice. On the basis of the size of the element and fluid velocity, a force is applied to each discrete element. Feedback from the element to the fluid model is achieved by requiring the fluid velocity on the solid element to be zero (no-slip boundary condition). A detailed development and validation of the coupled method can be found in work by Cook et al. [2000], Cook [2001], and Cook et al. [2004].

The two-dimensional simulations reported by Cook [2001] include such complex phenomena as drafting-kissing-tumbling in multiparticle sedimentation simulations and the saltation phase of bed erosion. Boutt et al. [2007] present an application of the model to problems in porous media, including transient flow, consolidation of a finite layer, and simulation of natural hydraulic fracture genesis.

5. Models of Natural Hydraulic Fracture Genesis

[31] Two-dimensional discrete element models were built using the LBDEM code to investigate the role of permeability and storage on the genesis of natural hydraulic fractures. Models were built using a space-filling algorithm that rains elements into a box, then packs them under gravity, subsequently confining them with boundary platens until a specified isotropic stress is reached. We present results with three different size models consisting of 1500, 3000, and 6000 elements, respectively. The smallest model (1500 elements) is $7 \times 10^{-2}$ m long and $3.5 \times 10^{-2}$ m high. The 3000 element model is $14 \times 10^{-2}$ m by $3 \times 10^{-2}$ m and the 6000 element model is $14 \times 10^{-2}$ m by $6 \times 10^{-2}$ m. Elements used in these simulations are ellipses, chosen because they result in frictional and mechanical properties more like those of natural rocks than models constructed of circular elements.

[34] To simulate the behavior of a cohesive rock, individual discrete elements are bonded to one another. Bonds are modeled as point-to-point constraints between neighboring particles using a simple spring formulation:

$$F_b = k_b \Delta x,$$

where $F_b$ is the force in the bond, $k_b$ is bond stiffness, and $\Delta x$ is relative displacement of the neighboring elements. Bonds are aligned with the surface normals of the neighboring elements and connect the closest surfaces of adjacent elements. If $F_b$ is greater than or equal to the bond strength $F_b^{crit}$, the constraint is removed and bonded elements are allowed to move freely. A limitation of this bonding approach is that it has no implicit shear strength. In practice, the bond has some finite shear strength since it is modeled as a surface-to-surface contact, and any offset (normal or tangential) is greater than or equal to the $F_b^{crit}$ will cause the bond to fail. The bonding scheme has no interaction with the flowing fluid. Properties of the discrete element portion of the simulations are presented in Table 2.

[35] The simulated LB fluid has a kinematic viscosity of $2 \times 10^{-6}$ m$^2$/s, similar to that of the fluid used in the deformation experiment. Simulations consisted of approximately $5 \times 10^5, 1 \times 10^6$, and $2 \times 10^6$ fluid nodes for the 1500, 3000, and 6000 element simulations, respectively. The fluid lattice, shown as the area extending beyond the particles in Figure 8, allows the element assembly to elongate during dilation. The samples depicted in Figure 8 were tested under condition of a fixed fluid pressure gradient and the resulting steady state discharge through the system was recorded. The bulk model permeability calculated from these measurements of the 3000 element simulation is $1.3 \times 10^{-8}$ m$^2$, a high value relative to cemented rocks, which range from $10^{-9}$ to $10^{-19}$ m$^2$. This high bulk permeability results from the 2-D character of the model. Specific properties of the fluid lattice are given in Table 3. The permeability of the model is modified by reducing the diameter of particles (term herein the fluid radius) that are allowed to interact with the fluid. For a more complete discussion of this parameter and its effect on permeability/porosity of the models see Boutt et al. [2007].

[36] The initial conditions described for the fracturing experiment were applied to the LBDEM model (Figure 8). Because the lattice-Boltzmann method used here is not stable at large differential pressures [Boutt et al., 2007], we must scale element properties and boundary conditions to achieve a numerically stable solution. Therefore, the properties listed in Table 2 (namely the normal and shear stiffness and bond properties) and the boundary conditions and initial pressure of the model are all scaled downward by 6 orders of magnitude. All mechanical properties still maintain the same relative relationship to one another.

### Table 2. Parameters of Solid Assembly

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Friction</td>
<td>0.5</td>
</tr>
<tr>
<td>Element density (kg/m$^3$)</td>
<td>2600</td>
</tr>
<tr>
<td>Element normal stiffness (N/m)</td>
<td>100</td>
</tr>
<tr>
<td>Element shear stiffness (N/m)</td>
<td>600</td>
</tr>
<tr>
<td>Element size range (m)</td>
<td>$1.1 \times 10^{-3} - 0.7 \times 10^{-3}$</td>
</tr>
<tr>
<td>Bond strength, $F_b^{crit}$ (N)</td>
<td>$7 \times 10^{-4}$</td>
</tr>
<tr>
<td>Bond stiffness, $k_b$ (N/m)</td>
<td>100</td>
</tr>
</tbody>
</table>
Biaxial tests performed on discrete element models (i.e., no fluids) with unscaled and scaled properties yield identical normalized results.

[37] Initially the model area is loaded isotropically with boundary platens (long rectangular shaped elements in which an applied force is maintained over a fixed area) to an effective stress of 5 Pa, then the effective axial stress (sample short axis) was dropped to 4 Pa while holding the lateral stress (sample long axis) constant. The lateral stress is maintained via platens while the axial stress is applied by maintaining a constant confinement force to discrete elements along the short axis of the model. This allows both the application of an axial load and movement of fluid through the boundary elements. The initial fluid pressure within the sample is constant. The bond strength of the elements is $7 \times 10^{-4}$ N for all models. This value is reasonable when compared to the magnitude of the initial fluid pressure drop across the sample, 17 Pa, which we refer to as an intermediate value of fluid overpressure. In scaled terms, the bond strength is akin to a tensile strength of 4 MPa for a fluid pressure drop of ~17 MPa across the sample. Element properties are consistent with those of the Abelsonstone.

[38] The model test begins with a simultaneous drop in the minimum principal stress and the fluid pressure at the boundaries. For all tests, the axial stress is dropped to 0.01 Pa and the fluid pressure (termed herein fluid overpressure) difference for the 3000 element simulation ($P_{f0} - P_f$) is 17 Pa. Fluid pressure and velocity were monitored along a line parallel to the long axis of the model specimen. In addition, screen shots of all the simulations were taken at 0.001 s intervals. Model inelastic deformation is tracked through a visual record of bond breakage over the course of the model experiment. Zones in which contiguous bonds are broken are considered model fractures.

5.1. The 3000 Element Model Fracture Results

[39] The numerical models were executed for 2 s of model time, roughly the time required for the system to come to a new equilibrium state. Snapshots of early through late time bond breakage in an example 3000 element simulation are presented in Figure 9. As the axial stress is dropped on both boundaries, forces pushing individual elements together are reduced, resulting in extension of the elastic framework. The additional fluid pressure drop on the left and right boundaries causes large pressure gradients to form near (but not at) the ends of the discrete element assembly, and subsequently propagate inward, broadly consistent with the poroelastic model (compare to Figure 4). Bonds begin to break first near these sample ends. Individual bond breakages coalesce into larger, mesoscopic features that locally split the assembly roughly parallel to the maximum principal stress. Test runs of dry models with identical boundary and initial conditions were performed with no resulting fractures, although the model does experience dilation with less than 1% axial strain. Saturated models with no fluid pressure drop on the boundaries (i.e., drained conditions), but with a drop in the axial stress, also did not form fractures. This confirms that the fluid overpressure is the key factor causing fractures to form in these simulations; in other words, they are model hydraulic fractures. We infer

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kinematic viscosity $(\text{m}^2/\text{s})$</td>
<td>$2 \times 10^{-6}$</td>
</tr>
<tr>
<td>Relaxation time</td>
<td>0.625</td>
</tr>
<tr>
<td>Node spacing (m)</td>
<td>$9 \times 10^{-5}$</td>
</tr>
<tr>
<td>Fluid radius</td>
<td>0.65</td>
</tr>
<tr>
<td>Time step (s)</td>
<td>$1 \times 10^{-5}$</td>
</tr>
</tbody>
</table>
on the basis of the timing of formation of individual fractures relative to the position of the steepest pressure gradient that either the gradient itself plays a role in breaking element bonds, or the bonds are most likely to break where pore fluid pressure is highest at any given time.

Both the large throughgoing fractures and smaller fractures produced in LBDEM models are roughly perpendicular to the minimum principal stress. This result is consistent with the orientations and magnitudes of the model stresses. Individual fractures prefer pathways where model elements are oriented with long axes radially at a high angle to the minimum principal stress, which is effectively broadside to the overall fluid pressure gradient. Local deviations in fracture orientation record preferential breakage along grain contacts in these apparently optimal orientations. It is important to note that the shape preferred orientation or mechanical anisotropy exhibited by the model as a whole (Figure 10) is distinct from the shape preferred orientation of grains along the fracture paths. This could be explained by the fact that (1) with one bond per contact length, bonds between the long sides of ellipses may be mechanically weak relative to bonds that connect elements end to end and/or that (2) such elements have more surface length exposed to prevailing fluid pressure gradients, resulting in larger net drag forces on the elements.

Within these models, fluid pressure gradients develop in response to reduced pressure at the boundaries and higher pressure within the system. The gradients appear to induce interparticle forces sufficient to break bonds, particularly (as noted above) at contacts between elements with specific orientations. After initial bond breakage, fluid pressure drops in a newly opened fracture and causes positive fluid pressure gradients toward the fracture (Figure 9). Despite this local forcing, macroscale fluid pressure gradients are large enough to allow the fracture walls to move apart in response to reduced boundary constraints and fluid-induced seepage forces. Local fluid forces are apparently not high enough to cause bond breakage, suggesting that local-scale pressure gradients are not important in fracture propagation at the macroscale. Alternatively, fractures form where fluid forces act in concert with locally optimally oriented model elements.

### 5.2. Role of Rock Permeability

Within these models, fluid pressure gradients develop in response to reduced pressure at the boundaries and higher pressure within the system. The gradients appear to induce interparticle forces sufficient to break bonds, particularly (as noted above) at contacts between elements with specific orientations. After initial bond breakage, fluid pressure drops in a newly opened fracture and causes positive fluid pressure gradients toward the fracture (Figure 9). Despite this local forcing, macroscale fluid pressure gradients are large enough to allow the fracture walls to move apart in response to reduced boundary constraints and fluid-induced seepage forces. Local fluid forces are apparently not high enough to cause bond breakage, suggesting that local-scale pressure gradients are not important in fracture propagation at the macroscale. Alternatively, fractures form where fluid forces act in concert with locally optimally oriented model elements.

#### Figure 9

Time series showing fracture initiation and propagation under realistic laboratory boundary conditions in the 3000 element LBDEM model. A complex evolution of fractures in the simulation results from a strong feedback between fluid pressure and stress state. In this and all models presented in section 5.1, element color reflects percent of bonds broken, with orange/red indicating intact bonds and blue recording broken bonds. Color shading behind elements is fluid pressure.

#### Figure 10

Element long-axis orientations plotted on a rose diagram demonstrate a strong shape-preferred orientation with mean ellipse long-axis orientation \(\sim 30^\circ\) clockwise from horizontal. The smaller subset of elements with long axes \(\sim <30^\circ\) clockwise from vertical to vertical appears to control the local orientations of major fractures in the model simulations, resulting in stepped, subvertical fractures.
modified by varying the discrete element’s fluid radius. Changing the fluid radius does not change the solid (DEM) porosity, but it does change the amount of space available for fluid to reside. Changing the “fluid porosity” of the sample changes sample storage. Table 4 lists the properties of the models used for the permeability sensitivity study. Sample permeabilities were calculated using Darcy’s law with a steady state, constant pressure drop. Storage of the samples was evaluated using a transient technique described by Boutt et al. [2007]. Even though sample permeabilities vary by more than 2 orders of magnitude, storage shows less variance. Hydraulic diffusivities of the samples vary no more than 1 order of magnitude. The time constants for the samples are correspondingly different by an order of magnitude. Run times to reach mechanical equilibrium in the

<table>
<thead>
<tr>
<th>Identification Number</th>
<th>LB Radius</th>
<th>Fluid Porosity (%)</th>
<th>Permeability (m²)</th>
<th>Diffusivity (m² s⁻¹)</th>
<th>Storage (m⁻¹)</th>
<th>Time Constant (s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.6</td>
<td>62.6</td>
<td>1.2 × 10⁻⁸</td>
<td>23.5</td>
<td>2.7 × 10⁻⁸</td>
<td>41</td>
</tr>
<tr>
<td>2</td>
<td>0.725</td>
<td>49.3</td>
<td>7.0 × 10⁻⁹</td>
<td>15.5</td>
<td>9.0 × 10⁻⁷</td>
<td>62</td>
</tr>
<tr>
<td>3</td>
<td>0.8</td>
<td>40.4</td>
<td>2.4 × 10⁻⁹</td>
<td>5.4</td>
<td>6.2 × 10⁻⁷</td>
<td>178</td>
</tr>
<tr>
<td>4</td>
<td>0.95</td>
<td>19.5</td>
<td>1.0 × 10⁻¹⁰</td>
<td>1.9</td>
<td>2.7 × 10⁻⁷</td>
<td>506</td>
</tr>
</tbody>
</table>

Figure 11. Results of fracture simulations for $k = 1.2 \times 10⁻⁸$ m², $k = 7 \times 10⁻⁹$ m², $k = 2.39 \times 10⁻⁹$ m², and $k = 1 \times 10⁻¹⁰$ m² permeability models. (left) Contoured fluid pressure results plotted on axes of time versus distance for a stationary horizontal line, with the horizontal line across the center of the each model domain. (middle) Screen shots of the model states at 0.2 s. (right) Filled contoured plots of fluid speed on axes of time versus distance along the same centerline. Plots are scaled to facilitate comparison of models.
assumptions were different because of these different sample diffusivities.

[43] Figure 11 shows results of the four models with varying permeability. The top model in Figure 11 has a permeability of \(1.2 \times 10^{-6}\) m\(^2\), similar to the model shown in Figure 9, and is subsequently referred to as the base model. The additional three models are positioned in order of decreasing permeability. Each model screen shot is accompanied by plots of both pore fluid pressure and fluid speed over a 0.05 s interval of model evolution. Two main features of the base model are important. The first is the time delay in the response of the fluid pressure as a function of distance from the pressure boundary condition. This time delay is directly related to the hydrologic properties of the model. The slight change from a deep red hue to cherry red is a pressure drop associated with reduced stress on the boundary. In this model, most of the fluid pressure has decayed to the boundary pressure by 0.2 s time. This delay is important with respect to fracture propagation and initiation.

[44] The second important feature of the base model is the location and timing of fluid pressure drops with distance across a horizontal transect through the center of the model domain. These pressure drops correlate with the locations (and times of propagation across the transect) of hydraulic fractures. Fluid pressure within a given fracture drops in response to the creation of pore space faster than the rate of inward flow following fracture initiation; the pressure response indicates a perturbation of the flow system. A high-velocity zone that can be attributed to fluid flowing into a newly opened fracture is evident at a time following fracture initiation in the fluid speed plot.

[45] Fracture patterns and bond breakages at 0.2 s are different for the 4 models. The most obvious difference is in fracture apertures. The highest-permeability model exhibits the largest apertures, whereas the lowest-permeability model has the smallest. The number of bonds broken in a given model increases with decreasing hydraulic diffusivity. This is consistent with the fact that as diffusivity decreases, fluid pressure within the pore spaces is higher for a longer duration, providing more opportunity for fracture growth. The highest-permeability model has very localized deformation completely surrounding the main hydraulic fractures. As the hydraulic diffusivity of the model is decreased, deformation becomes less localized. In addition, fracture propagation is faster in the higher permeability than in the lower-permeability models.

[46] Similar observations of spatial variations in fluid pressure and speed can be made in all of the models shown in Figure 11. Differences between lower- and higher-permeability simulations are interpreted to be due to the fact that high pore fluid pressure is maintained for a longer period of time in the lower-permeability models. This result is intuitive from time constant calculations (cf. Table 4). The timing of fluid pressure drops is consistent in all models. The locations of these drops are different; they are correlated to the locations of the fractures that develop in individual models. Although the magnitudes of fluid speeds at early times are similar in the highest- and lowest-permeability models, the later time magnitudes are strikingly different, reflecting differences in fracture development that we infer to be ultimately related to differences in the permeabilities of the models.

[47] In this permeability sensitivity study, the fluid pressure drop across the model domain was the same in each model run. This implies that the initial fluid pressure gradient in the system was the same for each model. The force of the fluid on the solid has been shown to be directly proportional to the fluid pressure gradient [Mourges and Cobbold, 2003]. We contend that the differences in temporal evolution in fluid pressure, which varies with diffusivity, are responsible for the differences in modeled fracture patterns. Specifically, the high porosity and permeability model has fractures that traverse the model domain and little other deformation. The lowest-porosity model, in contrast, has only one throughgoing fracture. Damage is otherwise more evenly distributed. On the basis of this model result, we propose that fluid forcing on the solid framework plays a major part in the generation of these fractures.

[48] The initial model-scale pressure gradients are the same among the models, but even at initial stages, the pressure gradients at the multiple-grain scale are different. This is a function of pore throat size, which varies from the high permeability to the low-permeability model. Figure 11 shows that internal pressure gradients were much steeper throughout the history of the lower-permeability models than in the higher-permeability models. We hypothesize that these steeper gradients are responsible for the more distributed deformation in the lower-permeability models. This hypothesis is explored in the next section, where the response of fracture genesis to permeability is considered using larger models to investigate controls on fracture spacing.

5.3. Fracture Spacing and Model Deformation

[49] Hydraulic fracture simulations were executed with the 3000 and 6000 element models to investigate relationships among fluid flow properties, fracture spacing, and model deformation (defined by the percentage of bonds in a model that were broken in the course of a given simulation). Additionally, a simulation with double the fluid overpressure was executed to evaluate the effect of increased overpressure on fracture spacing. Figure 12 summarizes simulation results for the 3000 element models as a function of permeability and fluid overpressure. Using model domains double the length of the 1500 element models allows two important aspects of hydrologic controls on fracturing to be examined: the influence of fluid pressure gradients on deformation, and fracture spacing at a greater length scale. The trends in fracture spacing and deformation in the 3000 element models are similar to those in the 1500 element simulations. For the intermediate fluid overpressure, fracture spacing measured along a line through the model center decreases from 0.0186 m for highest to 0.0093 m for the lowest-permeability model (Figure 12 and Table 5). Fractures are also less well defined and decrease in aperture with reduced permeability, consistent with the results of the 1500 element simulations. The overall deformation in the sample increases dramatically with overpressure: fracture spacing is 0.0093 m for the highest permeability and 0.0057 m for the lowest-permeability simulation. Deformation is so extensive that it locally approaches the length scale of the mean element diameter (0.0011 m).
Previous workers have evaluated relationships between bed or mechanical layer thickness and throughgoing fracture spacing, calculating the spacing to thickness ratio ($S/T_f$) for different layers [Bai and Pollard, 2000b]. They explain $S/T_f$ ratios = 0.8 or <1.2 by pure extension, and suggest this range of values records fracture saturation. Larger ratios are interpreted to indicate that the system has not saturated. Layered sedimentary rocks with $S/T_f$ ratios as low as 0.5 can be explained by infilling, however Bai and Pollard [2000b] propose that the minimum ratio for fractures formed by infilling is 0.8. They suggest that lower ratios would require mechanisms other than pure extension, such as fluid overpressure.

Most of the fractures in our simulations do not cut the entire model domain. To compare fracture densities among different models, we have measured the spacing of fractures that cut a scan line through each model domain, rather than counting only throughgoing fractures. For our simulations at intermediate overpressure, this fracture spacing normalized by model thickness ranges from 0.62 to 0.31; it varies from 0.31 to 0.19 for high overpressure. Although our results cannot be directly compared with those of Bai and Pollard [2000b], they support the conclusion that fracture density can be increased through hydrofracture formation, and suggest the new hypothesis that fracture spacing will decrease with increasing overpressure.

Figure 13 summarizes percentage of damaged (broken) bonds as a function of rock permeability for all of the simulations, showing that the percentage of damaged bonds increases (i.e., deformation increases) with decreasing permeability. The 3000 element simulation with a much larger overpressure (0.34 Pa) produces the most deformation of all of the simulations and the closest fracture spacing (Table 4 and Figure 13). The 1500 element simulations, which are roughly half the length of the 3000 element simulations, exhibit greater deformation than either the 3000 or the 6000 element models, suggesting some dependence of simulated deformation on model length or fluid pressure gradients. In contrast, the 0.06 m thick 6000 element simulations show similar deformation and fracture spacing to the 0.3 m thick 3000 element simulations. Thus, model thickness does not influence fracture spacing, and the $S/T_f$ ratios are lower for

<table>
<thead>
<tr>
<th>Fractures</th>
<th>Mean Spacing (m)</th>
<th>Spacing/Model Thickness</th>
</tr>
</thead>
<tbody>
<tr>
<td>Intermediate Overpressure</td>
<td></td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>0.0186</td>
<td>0.62</td>
</tr>
<tr>
<td>11</td>
<td>0.0118</td>
<td>0.39</td>
</tr>
<tr>
<td>14</td>
<td>0.0093</td>
<td>0.31</td>
</tr>
<tr>
<td>High Overpressure</td>
<td></td>
<td></td>
</tr>
<tr>
<td>14</td>
<td>0.093</td>
<td>0.31</td>
</tr>
<tr>
<td>18</td>
<td>0.072</td>
<td>0.24</td>
</tr>
<tr>
<td>23</td>
<td>0.057</td>
<td>0.19</td>
</tr>
</tbody>
</table>

Figure 12. Screen shots and interpreted fracture patterns of (a–c) intermediate and (d–f) high overpressure simulations. Both number of fractures and amount of deformation increase with decreasing permeability. The spacing of fractures in the model domain center decreases with lower permeability. Simulations with higher fluid overpressure exhibit a stronger dependence of fracture density on permeability. These latter simulations also exhibit smaller fracture spacing and greater overall deformation than their intermediate pressure counterparts.
6000 element models, suggesting a strong hydrologic control on the spacing of the observed fractures.

6. Conclusions

Experimental and numerical analyses suggest that the magnitude of both permeability and fluid overpressure may significantly affect the development and the occurrence of natural hydraulic fractures in granular porous media. Theoretical and numerical arguments suggest that hydraulic diffusivity (controlling timing of fracturing and longevity of overpressures) and rock permeability, which govern fluid pressure gradients in rock, are important properties in both the generation and propagation of fractures.

We successfully produced a fluid pressure gradient in a laboratory experiment, with pore fluid pressure high enough to generate natural hydraulic fractures, in a low-permeability sandstone. Previous experimental analysis of the role of fluid pressure in the initiation and propagation of fractures has been limited to induced hydraulic fractures. Induced hydraulic fractures differ from natural hydraulic fractures in that the fluid pressure inside the fracture is always greater than the fluid pressure within the surrounding matrix. Imposing the conditions to form a natural hydraulic fracture is more difficult. Our test design relies on the relatively slow speed of fluid flow with respect to stress wave propagation. This test is the first to experimentally demonstrate that natural hydraulic fractures are possible. Two-dimensional axisymmetric poroelastic models indicate that the highest pore fluid pressure occurs near the boundary of the sample as a result of internal deformation of the sample following release of the axial load, explaining why the laboratory sample did not fail in the sample center.

It appears from poroelastic considerations that rock properties such as permeability and storativity may exert an important control on the location, time of formation, and spacing of natural hydraulic fractures. The experimental protocol described here can help test this hypothesis. Our ultimate goal is to probe the relative impact of mechanical versus hydrologic heterogeneities in natural hydraulic fracture genesis. A suite of experiments on rocks with varying permeability and storage will further constrain the role of hydrologic properties in fracture formation. In this way, we will better understand both how important hydrologic properties of rock are in controlling shallow crustal mechanical processes and the feedback loop between mechanical and hydrologic behavior of rocks in the shallow crust.

LBDEM numerical models were developed to extend interpretation of experimental results and to evaluate fracture genesis in materials with similar mechanical characteristics but different permeability. Although the initial fluid pressure gradients produced in these models were identical, local gradients created where initial fractures formed affected the time evolution of inelastic deformation in some models. The lower-permeability models showed more deformation and less overall development of the largest fractures, resulting in a smaller percentage of throughgoing fractures and smaller fracture apertures than higher-permeability models. The higher-permeability models were less sensitive to local fluid pressure gradients and the primary fractures propagated without hindrance. This suggests that local-scale pressure gradients are not important in driving these fractures to propagate at the macro-
scale. Simulations of different dimensions of model thickness and length show a strong correlation between permeability and the overall deformation in the models. Simulations with twice the model lengths (3000 versus 1500 element simulations) exhibit slightly different fracture spacings and sample deformation. Ratios of fracture spacing to model thickness \( S/T \) calculated for all of the models are within the ranges suggested by previous studies of fracturing in sedimentary layers where additional fracturing mechanisms (beyond extension) are required to explain fracture oversaturation [e.g., Bai and Pollard, 2000b]. Doubling the thickness of a model layer does not affect fracture spacing, thus thicker models exhibit even lower \( S/T \) ratios, suggesting that the closely spaced fractures observed in some sedimentary rocks [Price, 1966; Ladeira and Price, 1981; Huang and Angelier, 1989; Narr and Suppe, 1991; Gross, 1993; Becker and Gross, 1996; Gross et al., 1997] may record large overpressures rather than mechanical layer thickness.

These results are of great importance to understanding fracture development in sedimentary basins. Specifically, they address the space-time evolution of such systems because of the strong link between hydrological and mechanical processes. They also may provide insight into fault zone deformation processes, where abnormal fluid pressure may play a role in the mechanics of earthquakes and faulting. Consideration of the time dependence of fluid flow may aid in understanding the temporal and spatial evolution of fault and damage zones. In addition, problems of fracturing in hot and pressurized hydrothermal aquifers [Natale et al., 1998; McTigue, 1986; Merlani et al., 2001; Natale and Salusti, 1996] may be addressed by models similar to those we have developed.

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References


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