THE EVOLUTION OF FAULT STRENGTH, PERMEABILITY, AND ACOUSTIC PROPERTIES IN EXPERIMENTAL STUDIES FROM FAULT INITIATION THROUGH THE SEISMIC CYCLE

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by
Bryan Kaproth-Gerecht

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The dissertation of Bryan Kaproth-Gerecht was reviewed and approved* by the following:

Chris Marone  
Professor of Geosciences  
Dissertation Advisor  
Chair of Committee  
Head of the Graduate Program

Charles Ammon  
Professor of Geosciences

Derek Elsworth  
Professor of Energy and Mineral Engineering

Eliza Richardson Marone  
Assistant Professor of Geosciences

Demian Saffer  
Professor of Geosciences

*Signatures are on file in the Graduate School
Abstract

Within Earth’s crust, fault zones accommodate significant deformation and strain resulting from plate tectonics and other processes. Due to the hazards associated with fault slip, much work has been done to understand the factors controlling deformation style within these zones, which can range from quiescent aseismic slip to devastating earthquakes, such as the 2011 Mw9 Tohoku Oki earthquake. In particular, our understanding of processes like slow earthquakes and healing within fault zones remains unclear. Additionally, as fault zones develop they become highly differentiated from their parent material, as fault materials mix, break, rotate, and develop into fabrics. These changes, which vary with fault composition, chemistry, stress, and strain, can cause significant strength changes and permeability decrease. In particular, fault permeability can dictate regional fluid flow and may allow faults to act as petroleum traps and seals. Despite the importance of such faults, our understanding of their permeability evolution, especially in marine-sediment basins, is relatively poor.

In this dissertation, I investigated the evolution of fault zones as they initiate and proceed through the seismic cycle. In particular, I studied the origins of slow earthquake slip, the mechanisms controlling deformation band formation, and the evolution of fault fabric and permeability with fault zone development. This work was predominantly conducted on laboratory fault zones in a biaxial forcing apparatus under conditions appropriate for fault development in Earth’s upper crust. In chapter 1, I present the first laboratory observations of repetitive, slow stick-slip in fault zone materials (serpentine) and mechanical evidence for their origin. In particular, we document a transition from unstable to stable frictional behavior above a threshold velocity of ~10 µm/s. Additionally, these events are accompanied by precursory elastic wave speed reduction (2-21%) that begins up to 60 seconds before failure, perhaps suggesting a reliable earthquake predictor. In chapter 2, I investigate fault zone evolution through the seismic cycle and as it initiates, documented via elastic wave speed measurements. These experiments were conducted on halite under conditions where pressure-solution is operative, and they show the interplay of elastic wave speed measurements with porosity and fabric formation. Indeed, these observations point to a new technique for non-invasive fabric observation within laboratory and natural fault zones. In chapter 3 and chapter 4, I also discuss fault zone initiation and development for two specific cases: deformation bands and clay-rich marine-sediment faults.
Chapter 3 highlights deformation band formation through laboratory experiments, and shows that fault strengthening via shear-driven comminution is the likely mechanism limiting strain. I observe significant strengthening at low shear strains (e.g., $\gamma < 5$), and tie these observations directly to particle-size reduction. To accommodate fault-like strain, many deformation bands form within a given region, significantly limiting fluid flow, similar to standard faults. In chapter 4, I discuss the permeability evolution of faults under conditions appropriate for marine-sediment basins, like the Gulf of Mexico. In particular, the role of halite and clay within faults adjacent to salt domes and the possibility for multimechanism behavior, including brittle deformation at high strain rates and ductile deformation and pressure solution at slower rates is unclear. We found that that fault permeability can be reduced by up to 2-4 orders of magnitude with clay content, that small load ($\sigma_n < 6$ MPa) and small strain ($\gamma < 5$) can cause $< 1$ and $< 2$ orders of magnitude permeability reduction, respectively, and that halite is largely interchangeable with quartz for permeability.

This dissertation provides mechanical insight on a variety of fault deformation styles, as well as their implications. I document the first observations of slow earthquakes in the laboratory, evidence for their origins, and evidence for viable premonitory earthquake signals. I characterize fault fabric evolution leading up to and through the seismic cycle, and suggest a new tool for these observations. I provide significant evidence for the mechanism controlling deformation band formation and arrest. And I show how permeability may evolve within marine-sediment basin faults.
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**Figure 2-2.** Porosity (A) and seismogram (B) evolve dramatically under virgin compaction ($\sigma_c = 0.5$ and 15 MPa) and with applied shear load. Experiment p3635 was conducted on pure halite (<125 mm) in saturated NaCl brine. The sample was compacted for ∼1 hour at 0.5 and for another hour at 15 MPa. The fault layers were then sheared 10 mm (10 mm/s), held for 3,000s, and sheared again for 2 mm (10 mm/s). Pressure solution is the dominant compaction mechanism in these experiments, and is responsible for the large drop in porosity from 50% to 10% over the course a few hours. The dramatic decrease in P- and S-wave flight time throughout these experiments is due to a combination of velocity increase and layer thinning, which was taken into account for velocity calculations.

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pressure solution) is evident throughout these micrographs, noted as $W$, and in some cases the halite appears to flow with the clay fabric, noted as $P$. Panels C and F highlight sister experiments with similar mixture ratios, but contain smectite (montmorillonite) clay and illite shale, respectively. These mixtures are similar, but the matrix appears more granular with illite than smectite, owing to the contrast in particle size and some quartz content of the illite shale.

**Figure 4-14.** SEM micrograph from smectite dominated experiment, p3821 (54% smectite, 33% quartz, 12% halite). Like the other experiments, halite welds with other grains ($W$) and is in brittle contact with some quartz grains ($B$). P-shears developed throughout this material, typical of clay-rich materials. Additionally, halite grains flow with and weld to the clay fabric, and apparently do not locally disrupt the clay fabric, unlike quartz grains.
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“Where the senses fail us, reason must step in”

– Galileo Galilei

“Question with boldness even the existence of a God; because, if there be one, he must more approve of the homage of reason, than that of bindfolded fear”

– Thomas Jefferson
Introduction

Large portions of Earth’s crust are critically stressed and in a state near failure. Regional stresses and strains build over time with plate motion and thermal relaxation, and stresses are relieved by deformation and faulting. Stress release can occur rapidly along faults, generating earthquakes, or slowly, through aseismic slip or slow-slip events.

Faults are localized frictional surfaces that develop within country rock and evolve to accommodate regional strain [e.g., Brace and Byerlee, 1966; Scholz, 2002]. Faults can develop in many types of rock and sediment, but the style of deformation, whether by earthquakes or steady aseismic slip, depends on a number of variables such as material, stresses, and chemistry within the fault [e.g., Brace and Byerlee, 1966; Faulkner et al., 2010; Ikari et al., 2011]. Some materials, such as quartz and granite, tend to weaken dramatically as fault slip velocity increases. In some cases, fault strength decreases faster than tectonic stress can be relieved, generating a force imbalance. This force imbalance drives acceleration, causing an earthquake [Scholz, 1998; Scholz 2002]. Alternatively, some frictional materials strengthen with accumulated slip and/or increasing slip velocity [Ikari et al., 2009a], and such is the case for many types of clay. In these cases, referred to as velocity strengthening, a force imbalance does not occur, and deformation is inherently aseismic. These behaviors are controlled by the rate- and state-dependent friction behavior of the fault gouge and the elasticity of the rock acting on the fault [Scholz, 2008].

The potential for earthquake generation depends in part on the evolution of fault strength with slip. Early attempts to understand earthquakes considered simple friction evolution from static to dynamic values. Thus, if this evolution happened over a small slip distance, relative to the country-rock elasticity, an earthquake might occur. We now know that the evolution from ‘static’ to ‘dynamic’ friction is not a simple one, and indeed even these ‘static’ and ‘dynamic’ values vary. In particular, frictional strength can be highly dependent on slip speed (rate) and contact quality (state) within a particular material, and taken together a rate- and state-variable friction law has been developed. One form of this law, known as the Dieterich-Ruina slowness law, was described in Scholz [1998], following \( \tau = \sigma_n [\mu_0 + a \ln(V/V_0) + b \ln(V_0 \theta/D_C)] \), where \( \tau \) is the shear
stress, and $\sigma_n$ is the normal stress. Within the brackets, $V_0$ is the reference slip velocity, $V$ is the slip velocity, $\mu_0$ is friction at $V=V_0$, $a$ is the direct-effect friction change, and $b$ is the evolution affect friction change, which occurs over a critical slip distance, $D_c$. The state variable, $\theta$, evolves as $d\theta/d\tau = 1 - \theta V/D_c$. Thus at steady state $\tau = \sigma_n [\mu_0 + (a - b) \ln(V/V_0)]$. For earthquakes to initiate in a given material, $(a - b)$ must be negative, driving $\tau$ decrease on an accelerating fault.

Recent works have shown a new class of fault slip representing part of the spectrum between aseismic slip and normal fast earthquakes [e.g., Dragert et al., 2001; Ide et al., 2007; Peng and Gomberg, 2010]. Slow earthquakes are best observed with creepmeter and geodetic observations, but generally release too little seismic energy to be detected via traditional seismology methods [Sacks et al., 1978; Dragert et al., 2001; Ide et al., 2007; Peng and Gomberg, 2010]. These events can rupture large fault regions, reaching the equivalent of a M 8 earthquake, yet they occur with slip speed only as high as $\sim$ 10-100 $\mu$m/s [Rubin, 2008; Peng and Gomberg, 2010]. The origin of these phenomena is poorly understood however, as the mechanism limiting slip speed, yet allowing dynamic rupture, has not been observed. Chapter 1 presents the first laboratory observations of repetitive, slow stick-slip, analogous to slow earthquakes, as well as evidence for their origin. This work, conducted by Chris Marone and myself, is currently under review with Science and suggests that slow earthquakes nucleate like normal earthquakes, with velocity weakening behavior driving dynamic instability. However, we observe a transition to velocity-strengthening at slip speeds above $\sim$ 10 $\mu$m/s, quenching the fastest slip velocities and allowing long-duration fault slip at velocities less than 100’s of $\mu$m/s.

Fault zones, and the gouge material that they develop, evolve throughout the seismic cycle (e.g., interseismic quiescence leading up to fault rupture). During fault slip the fault gouge dilates, contacts between grains are disrupted, and grains roll, break, and slide frictionally along one another [e.g., Bos and Spiers, 2000; Rawling and Goodwin, 2003; Niemeijer et al., 2008; Sammis and Ben-Zion, 2008; Samuelson et al., 2009; Kaproth et al., 2010; Chapter 2]. These changes affect the elastic behavior of the fault, which can be observed through elastic wave-speed measurements [e.g., Nur et al., 1998;
Fortin et al., 2007; Mavko et al., 2009]. Some of these changes may occur prior to earthquakes, as preseismic slip is a widely documented phenomenon. In nature, earthquake premonitory phenomena include foreshocks [Jones and Molnar, 1976], increased tremor activity [Bouchon et al., in press], nucleation zone deformation [Sacks et al., 1978; Linde et al., 1988], hydrologic signals [Silver and Valette-Silver, 1992], and changes in elastic moduli within the fault zone and surrounding wall-rock [Scholz et al., 1972; Aggarwal et al., 1973; Niu et al., 2008; Schaff, 2012]. Niu et al. (2008) observed two distinct precursor sound velocity excursions, 10 hours before a M 3 earthquake and 2 hours before a M 1 earthquake, along the San Andreas Fault at Parkfield, CA. While the universality of these observations is unclear, they may be manifestations of preseismic fault slip [Scholz et al., 1972; Roy and Marone, 1996; Beeler and Lockner, 2003; Schaff, 2012; Bouchon et al., in press]. In particular, chapter 1 shows elastic wave-speed changes prior to earthquake rupture occurring in tandem with preseismic slip.

In chapter 2 we discuss the evolution of fault gouge with fault initiation and through the seismic cycle using elastic wave speed measurements, and this work, conducted by Chris Marone and myself, is currently in preparation for submission to The Journal of Geophysical Research – Solid Earth. To emulate long-term healing processes that might occur in natural faults, we conducted these experiments on halite, under conditions where pressure solution is operative. Recent seismology studies have highlighted earthquake damage and recovery in areas surrounding faults. These studies focus on spatiotemporal changes in elastic wave speeds through the seismic cycle, and tend to show coseismic velocity decrease and subsequent recovery with log time [Baisch and Bokelman, 2001; Vidale and Li, 2003; Rubinstein and Beroza, 2004; Schaff and Beroza, 2004; Li et al., 2006; Chao and Peng, 2009; Sawazaki et al. 2009; Nakata and Snieder, 2011; Minato et al., 2012]. While these data cannot provide estimates of fault strength evolution, owing to their off-fault focus, they do provide insight on the extent of in situ earthquake damage and the processes of recovery. Chapter 2 provides direct observations across a laboratory fault zone, which we use to assess fault strength evolution leading up to fault slip.
Faults initiate as fractures, but these newly weakened zones become preferred for future regional strain accommodation. With early slip many changes occur within the fault zone: grains roll and break [e.g., Rawling and Goodwin, 2003; Sammis and Ben-Zion, 2008; Kaproth et al., 2010], asymmetric grains may develop preferred orientations, and highly asymmetric grains will align and build fabric networks [Logan et al., 1992; Haines et al., 2009; Haines et al., 2013]. Chapter 2 discusses fault evolution as the fault zone matures and fabric develops. Through continuous ultrasonic measurements, including P- and S-wave velocities and amplitudes, we observe the initial damage and dilation within the fault zone, the development and rotation of Riedel shears, and eventual formation of a steady-state fabric. These observations point to a new technique for the observation of fabric developments within laboratory or natural faults.

Chapter 3 and chapter 4 also discuss fault zone initiation and development for two specific cases: deformation bands and clay-rich marine-sediment faults. Deformation bands are miniscule fault zones that develop into pervasive networks. Each of these structures accommodates up to ~5 shear strain, but many bands in a given zone (~100 m wide) can accommodate broad regional strain. Deformation bands tend to occur in arenite sands (sandstone or unlithified sand), and previous works have shown their general development, architecture, and geometry in detail [e.g., Aydin, 1978; Cashman and Cashman, 2000; Schultz and Balasko, 2003; Fossen et al., 2007]. Recent experiments have generated these features in the lab [Mair et al., 2000], however the mechanism controlling their formation but limiting their total slip remains unclear. Chapter 3 [Kaproth et al., 2010] discusses deformation band formation in laboratory experiments, and shows that fault strengthening via shear-driven comminution is the likely mechanism limiting strain. This work, conducted by Chris Marone, Sue Cashman, and myself was published in 2010 at The Journal of Geophysical Research – Solid Earth (See Kaproth et al., 2010). We observe significant strengthening at low shear strains (e.g., \( \gamma < 5 \)), and tie these observations directly to particle-size reduction. With strain hardening, deformation bands must arrest as they become stronger than the surrounding parent material. New strain preferentially occurs in neighboring, undeformed material, developing new deformation bands. Thus wide networks of deformation bands can develop, which can
have significant implications for fluid flow and petroleum reservoir quality. Standard faults can also play important roles for reservoir quality, and chapter 4 discusses this potential role for marine-sediment faults.

Developed faults within marine-sediment basins are critical geologic features for assessing fluid flow and petroleum trap quality. These features can reduce permeability up to five orders of magnitude relative to the country rock, especially as clay is mixed into the fault zone [Evans et al., 1997; Crawford et al., 2008; Ikari et al., 2009a]. Fault permeability is highly dependent on the fault-gouge composition, chemistry, and slip history. Especially with shear, clay within the fault zone can develop complicated networks of shear fabrics [e.g., Logan et al., 1992; Haines et al., 2013], orienting the clay grains, and developing highly anisotropic permeability conditions, restricting flow across the fault [Bos and Spiers, 2000; Bos et al., 2000a].

Chapter 4 highlights work conducted by Chris Marone, Sankar Muhuri, Marek Kacewicz, and myself, and discusses fault permeability of marine-sediment fault analogs, with different quartz/clay/halite compositions. These materials were unlithified, and experiments were carried out under low effective normal stress (6 MPa), appropriate for many Gulf of Mexico reservoirs.

The role of halite for fault permeability is particularly unclear, but may be important for reservoirs bounded by faults and salt domes, as halite may intermix into these fault zones.

This dissertation discusses changes within fault zones through the seismic cycle and as they initiate. With initial development we observe particle size reduction, grain rotation, and fabric generation, and each of these processes is dependent on variables like fault composition, stress state, and chemistry. These changes affect fault strength and permeability, which may impact future deformation and fluid flow through potential reservoirs. During the seismic cycle we observe repeated damage and healing, affecting the dynamic elastic moduli within the fault zone. Through ultrasonic measurements we can observe these processes and the potential role of preseismic slip for earthquake prediction. Finally, this dissertation presents the first observations for slow earthquakes in the laboratory and evidence for their origin. With a transition from weakening to strengthening at elevated slip velocities, these are the first results corroborating a mechanism for slow earthquake slip in nature.
Chapter 1: Slow earthquakes, preseismic velocity changes, and the origin of slow frictional stick slip

Abstract

Earthquakes have long been understood as frictional stick-slip instabilities, wherein stored elastic energy is released suddenly, driving catastrophic failure. Seismic rupture propagates at a few km/s, with fault slip speeds from 1-10 m/s, consistent with elastodynamic theory. However, tectonic faults also fail in slow earthquakes in which slip speeds are 10-100 µm/s and rupture durations extend over months to years. Slow earthquakes can rupture large fault regions, reaching the equivalent of M 8 or larger, yet their origin is poorly understood. Here, we present the first laboratory observations of repetitive, slow stick-slip in fault zone materials, and mechanical evidence for their origin. Our experiments were conducted on serpentine fault gouge under stresses appropriate for slow earthquakes in nature. The laboratory slow slip events are accompanied by precursory elastic wave speed reduction (2-21%) that begins up to 60 seconds before failure. We document a transition from unstable to stable frictional behavior above a threshold velocity of ~10 µm/s. Our data provide direct evidence for the hypothesis that slow earthquakes represent prematurely-arrested normal earthquakes. Slow tectonic fault slip can transfer stress to the seismogenic zone and slow earthquakes have been implicated in triggering damaging normal earthquakes. The connection to precursory changes in elastic wave speed may offer new insight in the search for reliable earthquake precursors.

Results and discussion

Slow earthquakes represent modes of fault slip behavior on a spectrum between steady aseismic slip and normal earthquakes [Scholz et al., 1972; Sacks et al., 1978; Dragert et al., 2001; Ide et al., 2007; Peng and Gomberg, 2010; Ghosh et al., 2012]. Like normal earthquakes, slow earthquakes can accommodate most of a fault’s slip budget, with equivalent magnitudes of 8 or larger; yet this slip occurs slowly, over days to months, rather than the few 10’s of seconds for normal earthquakes [Dragert et al., 2001;
Ide et al., 2007; Peng and Gomberg, 2010]. Slow earthquakes often occur adjacent to traditional seismogenic zones, and may load these earthquake-prone areas. Recent work suggests that slow earthquakes may abet potentially devastating earthquakes, such as the 2011 Mw 9 Tohoku Oki earthquake [Kato et al., 2012].

While observations of slow earthquakes abound in recent studies, the underlying processes that produce these self-sustaining, quasi-dynamic ruptures remain poorly understood [Roy and Marone, 1996; Ide et al., 2007; Peng and Gomberg, 2010]. One possibility is that slow earthquakes represent prematurely arrested normal earthquakes with slip-speed limited by poromechanical effects such as dilatant hardening or a transition from unstable, velocity-weakenening frictional behavior to velocity strengthening with increasing slip speed. While several mechanisms have been proposed [Rubin, 2008], direct links to slow earthquakes remain elusive. If slow earthquakes initiate like normal earthquakes, they may similarly exhibit precursory behaviors such as accelerating fault slip or elastic wave-speed reduction.

Here we describe laboratory observations of fault zone materials showing repetitive, slow stick-slip friction events that mimic slow earthquakes (Fig. 1-1). Each experiment includes 50+ stick-slip events with durations ranging up to 35 seconds (Fig. 1-2). Our suite of experiments shows that the laboratory fault zone undergoes a transition from velocity-weakening friction to velocity-strengthening behavior above slip rates of ~10 µm/s (Fig. 1-3). During each slip event the fault zone shows large changes in elastic wave speed (2-21% decrease), with precursory changes (1-3%) starting up to 60 s before slip failure (Fig. 1-4).

Our friction experiments were conducted on layers of powdered serpentine in the double direct shear configuration, using a servocontrolled, stiff biaxial load frame [Marone, 1998]. To promote stick-slip failure we reduced the shear-loading stiffness with an elastic element (k = 0.31 kN/mm; Fig. 1-1A, inset). The experiments include continuous, active elastic wave speed measurements across the fault layers using a piezoelectric source and receiver (1.4 MHz, 900 V). P-wave flight times, or time of flight from source to receiver, were determined by cross-correlation (Fig. 1-5) of the P-wave first arrival and early coda (Tfl) and serve as proxies for P-wave velocity through the fault.
layers. Experiments were conducted under constant shear velocity (0.5 to 30 µm/s) and normal stress ($\sigma_n' = 1$ MPa), plausibly appropriate for slow earthquakes in nature [Rubin, 2008; Peng and Gomberg, 2010].

Like natural earthquakes, fault slip velocity during these stick-slip events exceeded the imposed far-field velocity. Figure 1A gives the full record for one experiment, showing the character of stress drops and the corresponding stair-step pattern of fault displacement. Details of individual events (Fig. 1-1B) show 1-30 second stick-slip durations, average slip rates < 50 µm/s, and an evolution from impulsive small events to larger slow events with increasing fault displacement. These events resemble those from previous experiments conducted on halite [Voisin et al., 2008], and are distinguished by their repetitive nature. These slow stick-slip events were reproduced under similar experimental conditions at 0.5, 1, and 30 µm/s loading velocity (Fig. 1-2). For our suite of experiments, slip durations ranged up to 35 seconds, with 0.01-0.9 mm displacements and average slip velocities from 15-280 µm/s (Fig. 1-2). Stick-slip events generally released a few 10’s of kPa shear stress, roughly 5-10% of the frictional strength (Fig. 1-1a). Maximum slip velocities ranged from 60-1,300 µm/s, but peak velocities were generally sustained for < 1 s, with acceleration and deceleration periods lasting upward of 10 seconds (Fig. 1-2B).

To investigate the processes responsible for slow-slip events, we conducted additional experiments under stiff loading conditions and performed slide-hold-slide (SHS) tests (Fig. 1-6) and velocity-step tests (Fig. 1-7). Figure 3 summarizes results from these experiments, which are consistent with limited existing data [Ikari et al., 2011]. We determined the friction rate parameter ($a-b = \Delta \mu / \Delta \ln V$) and critical slip distance ($D_c$) using standard techniques [Marone, 1998]. In general $a-b$ decreased and $D_c$ increased with shear strain and hold time (Fig. 1-6A). The SHS test assessed friction behavior at slip velocities of ~0.01 to 1 µm/s and yielded ($a-b$) values from -0.198 to -0.002, $D_c$ from 5-105 µm, and $k_c$ from 2.1-11.4 kN/mm (Fig. 1-6A), exceeding the elastic stiffness of the loading system. Velocity-step tests assessed velocities from 3 to 1,000 µm/s and yielded $a-b$ values from -0.004 to +0.018. Figure 3 shows that serpentine is velocity weakening
at low slip velocities and that it is velocity strengthening above a threshold velocity of ~10 µm/s (Figs. 3 & S3).

Our experiments also included active-source ultrasonic measurements of elastic wave properties determined using piezoelectric transducers fixed within the sample assembly (Fig 1). P-wave flight time increased during stick-slip events, with precursory excursions starting up to 60 s before each failure (Fig. 1-4). Figure 4A shows $T_{fl}$ changes for a complete experiment, determined from early P-wave coda, along with the fault layer thickness and shear stress. The detailed flight time changes for each stick-slip event (Fig. 1-4B) are superimposed on long-term trends (Fig. 1-4A) that derive from changes in grain contact stiffness and coordination number, shear fabric development and layer thinning [Marone, 1998; Haines et al., 2013; Knuth et al., 2013]. Correlation coefficients and pick fidelity decreased late in the experiment (Fig. 1-4A) due to increased sample attenuation, and thus smaller elastic wave amplitudes (Fig. 1-5C). Figure 4B highlights nine stick-slip events, showing raw data, averages, and details of the changes prior to the slip event (inset to Fig 4B), measured from peak stress. With up to ~0.2 µs flight time increase, absolute $V_p$ through the fault layers decreased by 2-21% during these events. Following each slip event $T_{fl}$ recovered with log time, and in many cases the original flight time was recovered within 100 seconds (Fig. 1-4C). Prior to failure, $T_{fl}$ decreased by up to 0.06 µs, accounting for 10-25% of the total $V_p$ change during failure (Fig. 1-4B). These preseismic flight time changes occurred in tandem with, and were perhaps driven by, ultra-slow preseismic slip (Fig. 1-8).

Our measurements indicate that serpentine exhibits the necessary and sufficient conditions for dynamic earthquake rupture (i.e. $a-b < 0$ & $k < k_c$). Friction theory indicates that stick-slip initiates when shear stress ($\tau$) exceeds fault strength and the rate of friction weakening exceeds the elastic unloading stiffness. To generate a large, dynamic earthquake rupture, the conditions for instability must persist for a finite time as slip accelerates and the rupture expands. This requires continuous fault weakening as slip velocity increases (i.e. velocity weakening; Fig. 1-3, inset) [Scholz, 2002].

Our data indicate that stick-slip events initiate because serpentine is velocity weakening at low slip velocity. At low velocities, these $(a-b)$ values are very large
compared to previous works [Marone, 1998; Ikari et al., 2011]. However as slip accelerates, the serpentine fault gouge undergoes a transition to velocity strengthening frictional behavior \((a-b > 0);\) Figs. 3 & S3), limiting fault slip velocity. Previous works (e.g., Ikari et al., 2011) show \((a-b)\) dependence on slip speed, and perhaps owe to changes in steady state grain contact details (e.g., Marone, 1998). The long stick-slip deceleration periods (Fig. 1-1B) suggest that slip velocity slowly decreases as \(\tau\) is relieved, with fault strength adjusting to match \(\tau\) (Fig. 1-3, inset). Friction theory indicates that fault slip will arrest after \(\tau\) becomes smaller than the minimum fault strength. This means that more negative values of \(a-b\) will tend to generate larger stress drop. Consistent with this, we observe larger events at higher strains and with longer interseismic times (Fig. 1-1), both of which are associated with more negative values of \(a-b\) (Figs. 3, S2, S3).

To the best of our knowledge, our laboratory observations of repetitive, slow slick-slip in fault zone materials, coupled with corroborating measurements of friction constitutive behavior are the first of their kind. We document the properties of slow slip events and find a clear transition from velocity-weakening to velocity-strengthening behavior (Fig. 1-3). These data provide direct evidence for the hypothesis of rate-dependent friction for slow earthquakes and slow-slip events [Ide et al., 2007; Rubin, 2008]. While this work does not represent the natural range of temperatures, saturation states, pressures, and chemistry (e.g., Dragert et al., 2001; Peng and Gomberg, 2010), it indicates that slow earthquakes could occur in fault zones that transition from velocity weakening to strengthening during fault slip. Unlike previous examples of slow laboratory stick-slip with long durations [Voisin et al., 2008], this mechanism does not require a limited fault stiffness range, and may thus be applicable to a wide range of fault conditions [Rubin, 2008]. Additionally, we note that our results do not preclude other mechanisms for slow slip, including dilatancy hardening or small regions of unstable behavior surrounded by velocity-strengthening material [Rubin, 2008; Peng and Gomberg, 2010; Ghosh et al., 2012], wherein models generate rise times similarly shaped to our observations (long acceleration and deceleration periods), but with very short interseismic times [Rubin, 2008]. Our findings further suggest that slow slip can occur
over a continuum of fault velocities (e.g., Peng and Gomberg, 2010) (Fig. 1-2), and thus moment-release rates, dictated primarily by the transition between rate-weakening and rate-strengthening behavior (Fig. 1-3), which may vary greatly with the chemistry, pressure, and composition of a given fault zone, like other friction parameters (e.g., Marone, 1998).

Earthquakes involve fracturing and frictional failure, which commonly affect the elastic properties within the fault zone and surrounding rock [Niu et al., 2008; Rivet et al., 2011; Schaff, 2012]. Friction data and instability theory indicate that quasi-static slip within the fault zone should initiate prior to earthquake nucleation [Dieterich, 1978; Beeler, 2003; Bouchon et al., 2013]. One expects that this preseismic deformation will produce changes in fault properties such as elastic wave speed; however, there are relatively few reliable observations of precursory changes in elastic wave speed prior to earthquake rupture [Niu et al., 2008; Schaff, 2012]. Perhaps a key limitation to these observations is proximity to the fault zone, as many of these studies focus on the broader region surrounding faults [Rivet et al., 2011; Schaff, 2012].

Niu et al. (2008) studied elastic wave speeds immediately adjacent to the San Andreas Fault, and they observed coseismic changes with apparent precursors 10 hours before a M3 earthquake and 2 hours before a M1 earthquake [Niu et al., 2008]. Our study also benefits from close proximity to the fault zone, and we observe very large $V_p$ changes (up to 20%), along with precursors that account for 10-25% of these changes, which start up to 60 seconds before failure (Fig. 1-4). After failure, we also observe velocity recovery with log time (Fig. 1-4C), consistent with field observations [Rivet et al., 2011] and extensive laboratory studies of frictional healing [Scholz et al., 1972; Dieterich and Kilgore, 1994; Marone, 1998]. Our precursory $V_p$ changes are most likely tied to preseismic slip and nucleation of instability, as indicated by precise measurements of fault slip (Fig. 1-8). Initial slip likely reduces $V_p$ by a combination of dilation [Marone, 1998; Fortin et al., 2007; Knuth et al., 2013], damage to partially healed grain contacts [Goddard, 1990; Nur et al., 1998; Fortin et al., 2007], and grain fracture [Fortin et al., 2007]. Our sound velocity observations are consistent with granular dilation and changes to stress state, grain coordination number, and grain contact quality [Goddard, 1990; Nur
et al., 1998; Fortin et al., 2007; Schubnel et al., 2007; Knuth et al., 2013], which evolve with shear deformation [Marone, 1998; Schubnel et al., 2007; Knuth et al., 2013].

Our experiments show that repetitive, slow stick-slip may be explained by changes in frictional properties as slip accelerates during dynamic rupture nucleation. Existing studies of rupture nucleation [Ben-David et al., 2010a, 2010b; Schubnel et al., 2011] show that variations in initial stress and fault strength can cause a broad range of dynamic rupture propagation rates and nucleation times [Ellsworth and Beroza, 1995]. Our observations of slow slip provide an additional mechanism for slow rupture propagation, one that can also account for reproducible, periodic events. We find that serpentine fault gouge undergoes a transition in frictional behavior at slip velocities of \( \sim 10 \mu \text{m/s} \), producing slip rates within the range for slow earthquakes. Moreover, we observe precursory \( V_p \) changes coincident with slip onset prior to stick-slip failure. These observations suggest: 1) that a renewed focus on field studies of temporal variations in elastic wave speed may be warranted in regions of high seismic hazard and 2) that a combination of slow slip monitoring and elastic wave speed measurement may provide a basis for reliable earthquake prediction in regions where damaging, normal earthquakes are triggered by slow fault slip.

**Supplementary methods and materials**

These experiments were conducted in a biaxial forcing apparatus under double-direct shear. Our configuration consisted of two fault zones sandwiched between three steel forcing blocks (Fig. 1-1A, inset), with normal stress applied horizontally and shear stress applied vertically. Fast acting servo-hydraulic controllers maintain specified conditions of constant horizontal force and constant vertical displacement rate. We measure force using strain gauge load cells, accurate to +/- 0.1 kPa for our sample dimensions (5 cm x 5 cm frictional contact area), and displacement using direct current displacement transducers, accurate to +/- 0.1 \( \mu \text{m} \). Stresses and displacements were recorded digitally at 10 kHz with a 24 bit system, and were averaged to 10-100 Hz for storage.
These experiments were run on serpentine (Eden Hills, VT) purchased from Wards. XRD analysis shows that this serpentine is predominantly lizardite with some antigorite, and contains <3% chlorite and pumpellyite. We powdered the serpentine rock in a rotary mill and sieved the <125 mm fraction, which was used in these experiments.

Experiments were carried out under constant normal load ($\sigma_n = 1$ MPa) on powdered serpentine, with 3 mm initial layer thickness, and at room temperature and humidity. We conducted three types of experiments. (1) Stick-slip experiments were carried out with a spring in the vertical axis to reduce the system stiffness. Sliding velocity for these experiments was held constant in the range 1 to 30 $\mu$m/s, with faster velocities used during initial loading (Fig. 1-1A). (2) The slide-hold-slide experiment (p3788) was conducted under stiff loading conditions (i.e. no spring). In total, three slide-hold-slide sequences were conducted as follows: shear 1 mm, 1 s hold, shear 1 mm, 3 s hold, shear 1 mm, 10 s hold, shear 1.5 mm, 30 s hold, shear 1.5 mm, 100 s hold, shear 1.5 mm. All shear was conducted at 10 mm/s shear loading velocity. The first, second, and third slide-hold-slide sequences were started at 3, 19, and 26 mm of net shear displacement, respectively. Each slide-hold-slide sequence was modeled using the Ruina rate/state friction law to determine best-fit frictional constitutive properties (i.e. $a-b$, $D_c$, $k_c$). (3) The velocity-stepping experiment (p3975) was also conducted under stiff loading conditions. In total, three velocity-stepping sequences were conducted as follows: 0.4 $\mu$m of shear in sequence from 10, 3, 10, 30, 100, 300, and 1,000 $\mu$m/s. The first, second, and third velocity-stepping sequence were started at 2, 16, 25 mm of net shear displacement, respectively. Each velocity step was modeled using the Ruina law and known elastic interaction for our apparatus to determine $a-b$.

Elastic properties were measured with two, 0.5” dual mode transducers (Lithium Niobate) purchased from Boston Piezo Optics. These transducers generate 1,400 kHz and 900 kHz P- and S-waves, respectively. We excited the right-hand transducer (Fig. 1-1) with a 900 V pulse with 0.1-4 $\mu$s pulse width generated by an Olympus NDT Model 5058PR pulser-receiver. At this voltage, the transducer generates a ~ 19 nm and ~ 1 nm amplitude P- and S-wave, respectively (compressional wave piezoelectric constant = 20.8 pm/V; shear wave piezoelectric constant = 0.8 pm/V). An identical transducer in the left
side block received the waveforms. We recorded the output response with a GaGe CS8382 multi-channel digitizer at 25 MHz (14 bit over a 1 V range). We focused primarily on P-waves throughout these experiments, because S-waves were emergent relative to the strong P-wave coda (Fig. 1-5). The signal to noise ratio was enhanced in this system by (1) using a very large excitation voltage, (2) stacking the waveforms (generally 20 per seismogram), and (3) shielding the transducers with Faraday cages.

We picked the P-wave and coda arrivals via cross correlation using standard seismic techniques. Cross correlation compares a master waveform to other seismograms, identifying a flight time shift. After selecting a master wave with high signal to noise ratio, we compared it against a moving window for every other seismogram, and made a pick at the highest correlation coefficient. One benefit of cross correlation is that it can offer 10x resolution compared to the sampling rate. Thus, we subsampled each master wave and seismogram with a 10x spline fit.

We calculated the absolute P-wave velocity throughout each experiment, following \( V_p = h(T_p - T^0_p) \), where \( h \) is the layer thickness, \( T_p \) is the P-wave flight time and \( T^0_p \) is the P-wave flight time from calibrations with no gouge layer. The calibration was carried out repeatedly to verify results: \( T^0_p = 11.36 \) µs. P-wave arrivals were noisy and thus we focused on velocity changes indicated by the P-wave coda. In general, the coda lagged behind the P-wave arrival with remarkable consistency (Fig. 1-1B), following it by \( \approx 2.4 \) µs. Therefore this term was used to obtain changes in absolute value of the P-wave velocity.
Figure 1. A full experiment showing repetitive, slow stick-slip. (A) Shear stress ($\tau$; black line) during repeating stick-slip failure of serpentine under constant far-field velocity (dashed orange line). Like earthquake cycles, $\tau$ builds over long periods but is released in punctuated slip events, creating steps in fault slip displacement (red line). Inset shows the double-direct shear arrangement and ultrasonic experimental setup. (B) Subplots show 30-second windows for each stick-slip event in A, with event durations from 1-30 seconds. Fault slip (gray lines) is driven by stress release (black lines). Note the clear progression from small to large events with increasing net displacement.
Figure 1-2. (A) Stick-slip duration and fault displacement across three experiments. Repetitive slow stick-slip occurred over a range of driving velocities (0.5, 1, and 30 µm/s) under otherwise identical conditions. Average slip velocities ($V_a$) typically ranged from 15-50 µm/s (full range from 15-280 µm/s). (B) Details of slip acceleration history for representative stick-slip events. Maximum slip velocities ($V_m$) typically ranged from 60-1,100 µm/s, but the fastest velocities were only sustained for short periods of time (<1 s). Acceleration and deceleration periods often lasted 10+ seconds, resulting in very low average velocities.
Figure 1-3. Serpentine’s strength dependence on fault-slip velocity. Serpentine is velocity weakening (−a−b) at low velocities (< ~10 µm/s), a necessary condition for earthquake initiation. At higher velocities serpentine becomes velocity strengthening, likely retarding fast slip. This behavior corroborates widely held theory for slow earthquakes (Ide et al., 2007; Rubin, 2008). These a-b values were determined from slide-hold-slide tests (SHS; Fig. S2) and velocity stepping tests (V-steps; Fig. S3), with initial and final imposed slip velocities shown on the x-axis. The inset shows a theoretical model for coseismic slip history. Fault slip begins when shear stress (τ0) exceeds fault strength, and slip accelerates so long as shear stress remains greater than fault strength. With coseismic stress release to lower levels (e.g., τf), fault slip decelerates above a threshold velocity, and accelerates below that velocity; thus the threshold velocity is maintained. With continued coseismic stress release, slip velocity decreases to balance τ and fault strength. Fault slip stops after τ drops below the minimum fault strength, driving deceleration.
Figure 1-4. Preseismic and coseismic elastic wave-speed decrease and postseismic healing. (A) Shear stress, fault thickness and $T_f$ throughout experiment p3787. With each slip event $T_f$ increases and gradually recovers. Figure 4B highlights nine sequential events (identified by color dashes in A), chosen for low noise and long recurrence intervals. We stacked these events (black line), to enhance signal:noise. Following peak stress $T_f$ changes were large, accounting for up to 2-21% $V_p$, however 10-25% of this decrease occurred prior to failure. Velocity changes preceded the main shear failure by up to 60s (B, inset). (C) Following each slip event, $T_f$ recovered with log time, indicating fault healing after rupture [Sacks et al., 1978; Dieterich and Kilgore, 2004; Marone, 1998; Nakata and Snieder, 2011] – 75% of baseline velocity was generally recovered after ~ 100 seconds.
Figure 1-5. (A) Over the course of one experiment, we collected 30,000+ waveforms. These seismograms are shown sequentially in panel A, with the earliest at the top and latest at bottom. An example seismogram (record 10,000) is shown above this plot; note the emergent P-wave arrival around 14 µs. To measure the P-wave arrival (B), we cross-correlated each waveform throughout the experiment against a master waveform (A, smaller inset). We determined $V_p$ throughout the experiment (D) using layer thickness measurements and $T_p$, calibrating out the flight time through the forcing blocks. Since the P-wave was emergent, we also cross-correlated two cycles of coda (A, larger inset), starting one cycle after the P-wave arrival. In general, the overall trends of the P-wave arrival and coda were similar, but the coda had much less noise (B), reflected in the correlation coefficients (C). For most of this experiment, we collected one seismogram per second (stacking 20 records). However, we collected unstacked seismograms at much higher rates (20Hz) over two periods (A). While these periods gave higher sampling resolution, the arrival picks (B) are very noisy, owing to increased seismogram noise without a stack. Additionally, the seismic signal got very weak after record ~ 12,000, increasing the flight time noise, likely owing to increased attenuation within the fault layers.
Figure 1-6. We conducted slide-hold-slide tests in a stiff system to determine the frictional properties of serpentine. During this experiment five hold periods (1-100 s) were conducted at low, medium, and high strain, each followed by 1.0 to 1.5 mm of shear at 10 µm/s. All other variables were held constant (room temp and humidity, $\sigma_n = 1$ MPa). Examples of data from the slide-hold-slide sequences are shown in (B). We modeled each sequence (Panel C) to determine $a-b$, $D_c$, and $k_c$ as a function of hold time and strain (A). Under all conditions, serpentine was velocity weakening ($-a-b$), satisfying the necessary condition for earthquake slip. At low strains, $a-b$ and $D_c$ were roughly independent of hold time. At all other strains, $b-a$ and $D_c$ were large, and tended to increase with hold time. The critical stiffness ranged from 2 to 11 kN/mm and tended to be larger at higher strains. The measured values of $k_c$ are higher than the elastic loading stiffness (0.31 kN/mm), and thus the friction instability was met at low velocity.
Figure 1-7. Data from an experiment under stiff loading conditions to determine friction behavior at higher velocities (i.e. 3 to 1,000 µm/s). Inset defines the variables $a$ and $b$. Conditions for this experiment were identical to the slow stick-slip experiments (room temp, 100% humidity, $\sigma_n = 1$MPa). Panel A highlights shear stress during three velocity stepping sequences at low (gray line), medium (blue line), and high (black line) strain. Each sequence shows step changes in loading velocity, separated by 0.4 mm of shear, from 3, 10, 30, 100, 300 and 1,000 µm/s, starting from an initial velocity of 10 µm/s. (B) We modeled each velocity step to determine friction constitutive properties as a function of velocity and strain. In general, $a-b$ increased with velocity, and decreased with strain. At medium ($\gamma = 9-10$) and high strain ($\gamma = 17-18$), serpentine was velocity weakening below slip rates of ~ 10 µm/s.
Figure 1-8. (A) Before each stick-slip event, the fault zone accommodates small amounts of slip (<20 µm). (B) Within 60 seconds before failure, $T_p$ increases proportionately with preseismic slip. Preseismic slip may cause $V_p$ changes by shear driven dilation or damage to well-healed grain contacts, and similar processes may drive precursor velocity changes in nature [Dieterich, 1978; Rubin, 2008; Niu et al., 2008; Schaff, 2012].
Chapter 2: Elastic wave speed observations of coseismic damage and healing within laboratory fault zones

Abstract

Earthquakes are dynamic frictional failure events that occur cyclically on a given fault. Earthquake rupture involves fault weakening and the release of elastic strain energy, and these periods are followed by long interseismic times wherein fault strength recovers through a set of processes known collectively as fault healing. Fault healing has been most readily observed in the laboratory, however recent seismology studies on elastic wave speed changes in Earth’s crust have yielded particular insight on the coseismic damage and recovery processes in fault damage zones and surrounding areas. Velocity observations within fault zones may provide further insight on fault strength evolution, however, observations of the fault zone itself are outside the resolution of past seismology studies, which are damage-zone scale or broader. Here, we document elastic wave speed changes within laboratory fault zones through cycles of damage and healing during fault slip and subsequent quiescence. These fault analog experiments were carried out on granular halite under brine-saturated conditions, as an analog to crustal rocks in the seismogenic zone where pressure solution is a key mechanism of frictional healing. Our experiments emulate the seismic cycle via slide-hold-slide tests and periods of quasi-static fault slip. Porosity ($\phi$) tends to be the dominant control on elastic wave speeds and amplitudes. During healing periods, porosity decreases with log time, decreasing up to 40% over a few hours; ultrasonic measurements of elastic properties reflect these changes. Elastic wave speeds and amplitudes increase systematically with waiting time between periods of shear; for each percent $\phi$ loss during compaction, $V_P$ generally increases by 3%, $V_S$ by 2%, $A_P$ by 10%, and $A_S$ by 7%. Dilation occurs during fault shear, consistent with granular deformation theory, but predominantly occurs with initial fault slip. At higher accumulated strains, fabric formation dominates the ultrasonic signals, as grain-scale heterogeneities develop and rotate within the sample. Indeed, ultrasonic observations may provide critical links to in situ observations of fabric formation during fault slip, both in the laboratory and in nature. Our observations provide clues for fault
healing through time, as both fault strength and elastic wave speed changes depend on interseismic porosity loss. In these experiments we find that a 1% change in $V_p$ or $V_s$ results in a respective fault strength increase of 0.01 or 0.02 $\Delta \mu$. Additional laboratory work and advances in seismic techniques to monitor elastic wave speed changes through natural faults may provide critical information on fault strength evolution and seismic hazard.

**Introduction**

Earthquakes have long been understood as dynamic frictional failure events along fault zones. Failure initiates when tectonic stresses exceed the fault strength, releasing elastic strain energy seismically and through damage to the fault zone. Following rupture fault zones heal and tend to regain strength with the logarithm of time [Scholz, 2002]. During this interseismic period, tectonic strain slowly rebuilds, accumulating elastic strain energy until the next rupture [Brace and Byerlee, 1966].

Recent seismology studies have highlighted earthquake damage and recovery in areas surrounding faults. These studies focus on spatiotemporal changes in elastic wave speeds through the seismic cycle, typically showing coseismic velocity decrease and subsequent recovery with log time [Baisch and Bokelman, 2001; Vidale and Li, 2003; Rubinstein and Beroza, 2004; Schaff and Beroza, 2004; Li et al., 2006; Chao and Peng, 2009; Sawazaki et al. 2009; Nakata and Snieder, 2011; Minato et al., 2012]. While these data cannot provide estimates of fault strength evolution, owing to their off-fault focus, they do provide insight on the extent of *in situ* earthquake damage and the processes of recovery.

Elastic wave speed changes have been observed for earthquakes including the 1979 M 5.9 Coyote Lake [Poupinet et al., 1984], 1999 M 7.1 Hector Mine [Li et al., 2003], 1992 M 7.5 Landers [Li et al., 1998; Li and Vidale, 2001; Vidale and Li, 2003], 1984 Morgan Hill [Schaff and Beroza, 2004], 2000 Western-Tottori [Sawazaki et al., 2009], 2004 Mw 6.6 Mid Niigata [Wegler and Sens-Schönfelder, 2007], 2004 M 6 Parkfield [Rubinstein and Beroza, 2005; Li et al., 2006], 2008 Iwate-Miyag Nairiku [Yamada et al., 2009], and the 2011 M$_{w}$9 Tohoku-Oki [Nakata and Snieder, 2011;
Nakata and Snieder, 2012] events. These studies used a variety of sources, including repeating earthquakes [e.g., Rubinstein and Beroza, 2005], chemical explosions [e.g., Li et al., 1998], and the rapidly-advancing field of ambient seismology [e.g., Nakata and Snieder, 2011]. Li et al. [2006] used fault-guided waves to study fault healing and they were able to provide the tightest spatial resolution around the fault zone to date, ~ 200 m. These damage zone observations showed ~ 2.5% S-wave velocity decrease after the M 6.0 Parkfield earthquake, followed by ~ 1% recovery over three months. Velocity loss surrounding the fault is thought to occur by processes of crack opening, regional stress changes, and fluid migration [Li et al., 1998; Li et al., 2003; Vidale and Li, 2002; Li et al., 2006; Silver et al., 2007; Niu et al., 2008; Schaff, 2012]. Similarly, it has been suggested that country rock permeability may also experience coseismic increase and subsequent healing [Elkhoury et al., 2006]. Damage within the fault zone more likely occurs by processes of dilation, comminution, and fabric generation. While these changes tend to be difficult to observe in nature, owing to the small width of many fault zones [Faulkner, 2010], recent advances in borehole instrumentation near fault zones (e.g. SAFOD; Niu et al., 2008) may offer tighter spatial resolution in the future.

Temporal velocity changes of fault zones are likely to be strongly influenced by porosity evolution through the seismic cycle. Previous experimental works relate porosity to elastic wave speeds through compaction experiments on sediments and/or across suites of similar rocks at different compaction states [e.g., Hadley, 1976; Scott et al., 1994; Popp and Kern, 2001; Schubnel et al., 2003; Fortin et al., 2005; Fortin et al., 2007; Croizé et al., 2010; Kitajima and Saffer, 2012]. Elastic wave speeds tend to increase linearly with compaction in the range of typical rock porosities, $\phi < 30\text{-}40\%$ [Erickson and Jarrard, 2009; Mavko et al., 2009]. However these works do not focus on repeated compaction and dilation in a given rock, which occurs within faults over the seismic cycle [Mead, 1925; Faulkner et al., 2010], thus the role of hysteresis has not been established. Additionally, fault zones develop fabric and heterogeneities with shear [Logan et al., 1992, Haines et al., 2013], which may result in unexpected velocity effects relative to bulk $\phi$ [e.g., Knuth et al., 2013].
Here, we present detailed spatiotemporal elastic wave speed changes within a laboratory fault zone through the seismic cycle. These observations show dramatic velocity decrease with fault slip, subsequent velocity recovery with log time, and highlight the importance of fault porosity for these changes. Additionally, we show that elastic wave speed observations may provide a useful tool to monitor fault fabric formation in situ. These experiments were conducted in the double-direct shear configuration using the biaxial forcing apparatus at Penn State University. To emulate coseismic slip, the fault zone was sheared under quasi-static conditions, yielding high temporal resolution of fault damage during fault slip. These fault analog experiments were carried out on granular halite under brine-saturated conditions, as an analog to crustal rocks in the seismogenic zone, where pressure solution is a key mechanism of frictional healing.

**Methods and materials**

Our experiments were conducted on ground halite in the biaxial forcing apparatus at Penn State. Figure 2-1 shows the double-direct shear geometry, which consists of two fault zones sandwiched between three steel blocks. For details see Anthony and Marone [2005] and Kaproth et al. [2010]. Normal stress ($\sigma_n$) is applied horizontally and displacement is imposed at the center block, inducing shear within the fault gouge layers. Fast acting servo-hydraulic controllers maintain specified horizontal load and vertical displacement rate. Strain gauge load cells, accurate to +/- 0.1 kPa, measured vertical and horizontal load. Direct current displacement transducers, accurate to +/- 0.1 µm, measured vertical and horizontal displacement (shear displacement and layer thickness, respectively). Stresses and displacements were recorded digitally at 10 kHz with a 24 bit system, and were averaged to 10-100 Hz for storage.

Experiments were conducted on halite, which was ground in a rotary mill and sieved to <125 µm. We constructed initial sample layers that were 0.5 cm thick and 5 cm x 5 cm nominal frictional contact area using a leveling jig. Layers were loaded to 0.5 MPa $\sigma_n$ and soaked in salt-saturated brine for one hour after which normal stress was increased to 15 MPa. Figure 2-2 shows the simple effect of this procedure on $\phi$. Samples
were sheared 10 mm following this initial loadup. This ‘run-in’ period develops steady state fabric within the gouge layers [e.g., Marone, 1998; Kaproth et al., 2010; Niemeijer et al., 2010], and will be discussed in detail in the following sections. After the run-in, shear was stopped for a 1 to 3,000 second hold period, after which shear was resumed for 2 mm. See Table 1 for experiment details. In experiment p3636 we conducted multiple hold periods (1 to 3,000 s), each followed by 2 mm of shear.

At the end of each experiment, shear stress was removed, the brine was drained, and the final sample thickness was measured. The samples were recovered intact from the side-blocks after normal stress was removed. Once recovered, the samples were flushed with isobutyl alcohol, oven dried, and weighed.

Continuous velocity measurements

Continuous ultrasonic measurements were made via two, 0.5” shear wave piezoelectric transducers (500 kHz, lead-zirconate-titanate) from Boston Piezo Optics (D31 PZT-5a). We excited the right-hand transducer (Figure 2-1) with a 900 V pulse generated by an Olympus NDT Model 5058PR pulser-receiver. At this voltage, a ~ 0.15 μm amplitude S-wave is generated (0.171 nm/V piezoelectric constant). P-waves are generated simultaneously by mode conversion, but their amplitude tended to be only ~ 5% of the S-waves. Elastic waves were received by an identical PZT transducer in the left side block (Figure 2-1). We recorded the output response with a GaGe CS8382 multi-channel digitizer at 25 MHz (14 bit). Waveforms are stacked, 100 per sample, and we record one waveform every two seconds (Figure 2-2B). A typical experiment included 1000+ waveforms.

Our instrumentation was designed to optimize P- and S-wave arrival picking. With an S-wave transducer, the S-wave is very large compared to the P-wave coda (Figure 2-3), making it relatively easy to pick. Additionally, the large excitation voltage used made the P-wave much larger than the background electronic noise. We built Faraday cages around each transducer to further isolate electronic noise.

We picked the P- and S-wave arrivals via cross correlation using standard seismic techniques [e.g., Stein and Wysession, 2003]. Cross correlation compares a master
waveform to other seismograms, identifying a flight time shift. After selecting master waves with high signal to noise ratios (Figure 2-3A), we compared them against a moving window for every other seismogram in the experiment, making a pick at the highest correlation coefficient (Figure 2-3). In general, we have found that cross correlation is most effective when the starting and ending points are near zero (≤ 3% from zero). Figure 2-3B compares the P-wave template to a seismogram from 500 seconds earlier. The template wave is compared against a moving window of the wave of interest. The pick is made at the maximum correlation coefficient (e.g., 12.6 μs; Figure 2-3B inset). One benefit of cross correlation is that it can offer 10x resolution compared to the sampling rate. Thus we subsampled each master wave and seismogram with a 10x spline fit. Owing to the similarity of waveforms, maximum correlation coefficients are very high, generally > 0.98 for P-waves and > 0.99 for S-waves (Figure 2-4).

While using S-wave transducers to generate simultaneous P- and S-waves offers distinct advantages (e.g., identical travel path, simple set up, etc.), it can make the S-wave difficult to pick if it arrives in phase with the P-coda. Our experiments included a simple solution to this problem: as the layer thins and velocities change, the S-wave migrates through the P-wave coda and eventually became out of phase (Figure 2-2). Because cross correlation only requires one master waveform, we select this when the S-wave is out of phase with the P-coda, and use it to automatically pick all other S-wave arrivals.

We calculated the absolute P-wave velocity throughout each experiment, following \( V_p = \frac{h}{T_p - T^0_p} \), where \( h \) is the layer thickness, \( T_p \) is the P-wave flight time and \( T^0_p \) is the P-wave flight time from calibrations with no gouge layer under similar normal load. Similarly, we calculated the absolute S-wave velocity following \( V_s = \frac{h}{T_s - T^0_s} \), where \( T_s \) is the S-wave flight time and \( T^0_s \) is the S-wave flight time from calibrations. The calibrations were carried out repeatedly to verify results: \( T^0_p = 11.1 \) μs and \( T^0_s = 19.1 \) μs.

We measure P- and S-wave amplitude (\( A_p \) and \( A_s \)) in these experiments at the peak following the first arrival (Figure 2-3). With a constant source and thin layer, thus limiting radial dispersion and attenuation, the dominant control on amplitude is the
reflection coefficient, which is directly tied to impedance contrast between the sample and the forcing blocks [e.g., Stein and Wysession, 2003].

Ultrasonic measurement uncertainty

The two primary contributors to $V_P$ and $V_S$ measurement uncertainty are layer thickness and flight time. While changes in $h$ were tracked precisely throughout each experiment, initial thickness was measured by hand with calipers to a precision of ±0.1 mm, contributing to absolute error. Additionally, forcing blocks in the double direct shear configuration are subject to some tilting with shear strain, adding up to ±0.15 mm of error. While this error is ‘relative’, these changes occur gradually and should not affect interpretation of short-term trends. Uncertainty in the flight time measurement was ± 0.08 µs, or < 7% of the signal from p3635, and is derived from calibration uncertainty as well as the digital sampling rate. $V_P$ and $V_S$ error tended to be largest at the end of an experiment since $h$ decreases with strain. Figure 2-5 shows absolute $V_P$ and $V_S$ uncertainties for p3635, which were $< ~ 35\%$ and $< ~ 10\%$, respectively. Our center $V_P$ and $V_S$ values are slightly greater than previously reported values for halite [Lazarus, 1949; Mavko et al., 2009], but they fall in range when error is considered (Figure 2-5). Figure 2-6 compares porosity-velocity trends (p3635) against the work of Popp and Kern [1998]. Although their work was on natural halite with some anhydrite and polyhalite, our trends are in general agreement.

Electronic noise was the primary contributor to $A_P$ error, resulting in a cloud of datapoints (e.g., Figure 2-5). This noise did not affect the much larger $A_S$ signal, but $A_S$ was more sensitive to occasional power fluctuations (evident during the initial compaction in Figure 2-5). The major player for $A_S$ error, however, was convolution with the P-wave coda. In general, the P-wave coda was $\sim 10\%$ of the S-wave amplitude (Figures 2-1, 2-3C-D), adding up to ±10% error for $A_S$. However, this error is dependent on the gradually changing phase match (Figure 2-2). For example, in p3635 the S-wave arrival was generally out of phase with the P-wave coda (160-200°), limiting the relative error in p3635 to ± 3.5%.
Continuous porosity measurements

We measured porosity at the beginning and end of each experiment, and we used these values to estimate $\phi$ throughout each experiment. At these endpoints, the sample mass ($M$), $h$, and contact area ($A$) are known. We calculate bulk density, following $\rho_{\text{bulk}} = M/(h*A)$, and porosity, following $\phi = 1-(\rho_{\text{solids}}-\rho_{\text{bulk}})/\rho_{\text{solids}}$, where $\rho_{\text{solids}}$ is the density of halite (2.16 g/cm$^3$). To calculate $\phi$ throughout each experiment, we account for $h$ (which is measured continuously) and mass loss via geometric thinning [Scott et al., 1994].

Geometric thinning is a phenomenon of mass shedding that occurs in any mode of simple or pure shear except ring shear [Scott et al., 1994]. This mass loss must be accounted for to accurately determine $\phi$ throughout shearing. The standard model for geometric thinning is: $h*\delta x/2 = L*\delta h_{gt}$, where $h$ is layer thickness, $\delta x$ is the instantaneous shear displacement, $L$ is sample length, and $\delta h_{gt}$ is the incremental thickness change [Scott et al., 1994]. This equation suggests that mass is lost as a triangle with shear, but alternate shapes are possible, especially considering the effects of localization and distributed shear [e.g., Rathbun and Marone, 2010]. In the interest of high fidelity $\phi$ measurements, we calibrated the Scott et al. [1994] model for our samples.

To calibrate the mass loss model, we compared measured mass loss against mass loss estimates from 14 experiments (Table 1). To estimate mass loss at any time during the experiment, we used a forward model with conservation of mass. We estimate mass lost at a future time $t$, following $M' = \delta x'*h'*y'*\rho'_\text{bulk}$, where $y$ is the width (5 cm) and $\delta x'$ is the incremental displacement. Therefore sample mass and mass lost at any future timestep can be predicted, following $M'^{+1} = M' - M'_{\text{lost}}$, or, alternatively $\rho'^{+1}_{\text{bulk}}*h'^{+1}*x*y' = \rho'_\text{bulk}*h'*x*y - \delta x'*h'*y'*\rho'_\text{bulk}$. There is uncertainty in density at the beginning of the experiment, owing to some mass extrusion when the sample is loaded. However, we have good mass constraint at the end of each experiment. Therefore we run this model backwards, following $\rho'^{-1}_{\text{bulk}}*h'^{-1}*x*y = \rho'_\text{bulk}*h'*x*y + \delta x'*h'*y'*\rho'_\text{bulk}$. Figure 2-7 shows a comparison of predicted mass lost versus the actual mass lost for these experiments (Table 1), and shows that a rectangular mass loss model fits our data well. We validated the rectangular loss model by back-predicting the initial sample mass, and comparing it to
the known initial sample mass (Table 1). Therefore, we used a rectangular model to predict the sample mass and $\phi$ throughout these experiments (e.g., Figure 2-5a).

Since these are brine-saturated samples, we accounted for the mass of salt precipitated out of the brine after the sample was dried. The actual porosity is calculated at the end of the experiment, following $\rho_{\text{meas}} = \rho_{\text{solids}} \ast (1 - \phi_{\text{actual}}) + 0.32 \text{ g/cm}^3$, where $\rho_{\text{meas}}$ is the measured bulk density at the end of the experiment, $\phi_{\text{actual}}$ is the final porosity, and $0.32 \text{ g/cm}^3$ is the amount of salt per cm$^3$ of saturated brine.

The two primary contributors to $\phi$ error are layer thickness and mass measurement error. Like velocity error, these factors predominantly affect the absolute values, since they are zeropoints that are carried throughout each experiment. Layer thickness error was discussed in section 2.1-2. Mass error comes from imperfect sample recovery at the end of an experiment – we assume $\pm 5\%$ as a precaution. A second source of mass error comes from the geometric thinning model (Figure 2-7). To estimate this error we considered a range of shape factors, from $1.2$ to $0.9$, which encompasses the data from our geometric thinning calibration. Figure 2-5 shows representative $\phi$ error for one experiment, from $+11$ to $-3\%$ relative to the center value.

Results

Under simple compaction, porosity in our laboratory faults zones dropped from $\sim 50\%$ to $\sim 10\%$ over a few hours. Figure 2-4 highlights $\phi$ evolution throughout a simple compaction experiment (p3455), similar to the load up period in the other experiments. Within $4,000$ seconds the sample compacted with log time from $50\%$ to $30\%$ porosity ($\sigma_n = 0.5 \text{ MPa}$). With increased load ($\sigma_n = 15 \text{ MPa}$), the sample underwent a second bout of compaction from $30\%$ to $8\%$ $\phi$ within $10,000$ s. The ultrasonic signals also increased dramatically with log time at each new load (Figure 2-4). Over the course of this simple compaction experiment $V_P$ and $V_S$ doubled and $A_P$ and $A_S$ increased tenfold. These ultrasonic behaviors likely reflect porosity evolution, and we discuss this relationship in detail in the following sections.
In general, P- and S-waves had large amplitudes, yielding high fidelity P- and S-wave arrival picks. Early in experiments, however, $A_P$ tended to be small (0.4 mV) compared to the electronic noise ($\pm 0.2$ mV), which affected the pick quality (Figure 2-4). However, even these poor picks had reasonable correlation coefficients ($> 0.8$; maximum range from -1 to 1). At higher loads, $A_P$ increased and the correlation coefficients became much higher (e.g., 0.95 after 15 MPa $\sigma_n$). In general, S-wave correlation coefficients were high ($> 0.97$). This quality reflects the signal strength of $A_S$, which was generally 20x larger than $A_P$. All experiments generally had similar histories with early compaction (Figures 2-4 and 2-5). Figure 2-8 shows the full experiment history following initial compaction for p3635, an experiment that included shear.

**Shear experiments**

To investigate fault gouge evolution with fault slip and subsequent healing, we conducted experiments with periods of shear and quiescence, analogous to the earthquake cycle. The following sections focus on two ‘slide-hold-slide’ experiments, p3635 and p3636. Following initial shear, p3635 had one hold period (3,000 s), followed by 2 mm of shear (Figure 2-8). Whereas p3636 had multiple hold periods (1-3,000 s), each followed by 2 mm of shear (Figure 2-9). During these experiments, the fault zones compacted during holds and dilated with fault slip.

During shear samples tended to dilate initially, followed by steady gradual compaction with strain (Figures 2-8 and 2-9). Figure 2-10 highlights the deformation that occurs during initial shear, along with attendant changes to $V_P$, $V_S$, $A_P$, and $A_S$. With initial load the sample deformed elastically and $\tau$ increases quickly. Near peak $\tau$ the sample transitioned from elastic to anelastic deformation, and frictional deformation began. After this transition, $\tau$ peaked and the sample slipped at a residual frictional strength. Until ~ 3 mm of shear, the sample dilated dramatically, from 9% to 17% $\phi$. Dilation occurred with initial shear across all experiments and peaked within the first ~ 2 mm of shear (~ 40% shear strain). Following this, samples tended to slowly and steadily compact with strain. During shear the ultrasonic measurements tended to reflect these porosity changes.
With shear there were three primary phases of behavior for the ultrasonic measurements. Figure 2-10 highlights these changes for initial shear during p3635: 1) with the onset of load and dilation the $V_P$, $V_S$, $A_P$, and $A_S$ remained steady (until ‘a’; Figure 2-10), 2) through the period of rapid dilation these signals dropped off dramatically, to near pre-hold levels (from ‘a’ to ‘b’; Figure 2-10), 3) after peak dilatancy rate (peak $\delta\phi/\delta\gamma$), these trends inflected, and the similarities between velocities and amplitudes broke down (after ‘b’; Figure 2-10). For the remainder of the shear period following peak $\delta\phi/\delta\gamma$ (2 to 10 mm) $V_P$ and $V_S$ briefly increased (2-5 mm), decreased (5-9 mm), and increased again (9-10 mm). Alternatively, $A_P$ and $A_S$ abruptly decreased from 2-4 mm and slowly increased for the remainder of shear. Similar trends occurred with initial shear during p3636 (Figure 2-9), and in both experiments with shear following long holds (e.g., 3,000 s; Figures 2-8 and 2-9).

Experiments with chemically assisted healing via pressure solution such as these, tend to have very large peak stress values following holds. This reproducible peak is related to fault healing, and tends to increase with log time. The healing rates from p3635 and p3636 were 0.15 $\Delta\mu$ per decade ($\Delta\mu$ is in friction coefficient change, or $\Delta\tau/\sigma_n$), similar to previous experiments [Niemeijer et al., 2008].

Similar to compaction during initial loading, the samples compacted during holds following shear. Porosity decrease occurred with attendant increases in $V_P$, $V_S$, $A_P$, $A_S$, bulk modulus ($k$), and shear modulus ($G$; Figures 2-8 and 2-9). Figure 2-11 highlights these changes as a function of hold time. In this particular case, porosity decreased 26% $\Delta\phi$ per decade increase in hold time (e.g., 1, 10, 100s, etc.). In general, $A_P$ was the most sensitive parameter during holds, increasing by 125% during the 3,000s hold (Figure 2-11), and $V_S$ was the weakest, increasing by only 25%. Figure 2-12 shows these same data along with three other experiments (p3453, p3456, and p3457), highlighting experimental reproducibility. During holds following shear, $\tau$ decayed with log time, resulting from sample creep under load (e.g., Figure 2-8). This behavior is typical during holds in many granular materials, but the nearly complete shear stress decay over a short period of time is unique to halite (Figures 2-8 and 2-9).
While the bulk and shear moduli were sensitive to changes during these experiments (e.g., Figure 2-11), we do not focus on them in depth. These variables are derived, compounding error from $\rho$, $V_P$, and $V_S$ [Stein and Wysession, 2003]. In particular, note the high bulk modulus values in experiment p3636 (Figure 2-9), which far exceed the single crystal halite value, 24.8 GPa [Mavko et al., 2009].

Dilation with shear and compaction during holds appears to be robust. These trends continued regardless of total displacement (0-23 mm) and over many slide-hold-slide cycles in experiment p3636. In particular, the 3,000 s holds and subsequent shear periods from p3635 (Figure 2-8) and p3636 (Figure 2-9) appear to be very similar.

**Porosity-velocity relationships**

In general, porosity appears to control elastic wave speeds and amplitudes. Under simple compaction these signals clearly increased with porosity decrease (Figure 2-4), and in more complicated experiments, this relationship holds with compaction and dilation (Figure 2-6). Figures 2-13 and 2-14 compare $\phi$ against $V_P$, $V_S$, $A_P$, and $A_S$ for experiments p3635 and p3636, respectively. In the following section, we generally refer to the left-hand plots of figure 2-13 (initial compaction and initial shear periods of p3635), but these behaviors are repeated throughout these experiments.

During initial compaction $V_P$, $V_S$, $A_P$, and $A_S$ generally increased linearly with $\phi$ decrease (Figure 2-13; left panels, black lines). Across all of these experiments, per 1% $\phi$ decrease, $V_P$ generally increased by 3%, $V_S$ by 2%, $A_P$ by 10%, and $A_S$ by 7%. Similar trends occurred following holds throughout these experiments (Figure 2-11).

During shear, each of the ultrasonic signals generally decreased with dilation (Figure 2-13, gray lines). However, these signals broke from linear trends with $\phi$ after the maximum rate of dilation, or peak $\delta \phi / \delta \gamma$, was achieved (Figures 2-13 and 2-14). This initial shear period (0 - ~ 2 mm) accounts for a majority of the $V_P$, $V_S$, $A_P$, and $A_S$ decrease, and similar changes occurred following holds (Figure 2-13, right panels). Following peak $\delta \phi / \delta \gamma$ however, $V_P$ and $V_S$ started to scale positively with $\phi$. This relationship occurred in both p3635 and p3636, during the run-in and after both 3,000s holds. The amplitudes had a different behavior from velocities after the peak $\delta \phi / \delta \gamma$.
generally becoming more sensitive to dilation (Figures 2-13 and 2-14). These aberrant trends held until ~ 8 mm of shear, after which elastic wave speeds and amplitudes resumed their typical trends with bulk $\phi$.

After short holds, the velocities and amplitudes decreased with dilation, and showed little to no second order effects (Figure 2-14). In particular, the velocities experienced little hysteresis with respect to compaction during holds. The amplitudes, however, were more sensitive to porosity changes with dilation than compaction (Figure 2-14).

Through the healing periods, $\phi$ decreased rapidly and the elastic wave speeds directly reflected these changes. Compaction similarly caused strength increase within the samples, as peak strength is controlled by dilation during initial shear [Marone et al., 1990]. Figure 2-15 shows that the fault strength increases by ~ 0.15 $\Delta\mu$/dec. during holds, along with $V_p$ and $V_S$ increases of 15%/dec. and 7%/dec., respectively. Since these variables are commonly dependent on $\phi$, elastic wave speed changes might be a useful tool to monitor fault strength evolution in nature. Per percent change in $V_p$ and $V_S$, fault strength increased by ~ 0.01 and ~ 0.02, respectively (Figure 2-15C).

**Discussion**

These experiments replicate coseismic damage and subsequent healing within the fault zone and allow us to monitor fault gouge evolution through these periods. During fault healing, we observe ~ 20-40% elastic wave speed increase and ~ 50-100% amplitude increase, far exceeding velocity changes surrounding faults in nature (e.g., ~ 2% $V_S$ increase; Li et al., 2006). These changes result from dramatic compaction during interseismic periods following shear: for every 1% $\phi$ loss, $V_p$ generally increases by 3%, $V_S$ by 2%, $A_p$ by 10%, and $A_S$ by 7% (Figure 2-12). During fault shear we observe large elastic wave speed and amplitude decrease early, when dilation is dominant, followed by subtle velocity changes. This finding indicates that while elastic wave speed changes may be useful for healing estimates during interseismic periods, they may not be reliable to estimate coseismic fault damage.
The role of compaction

Healing during these experiments occurred via pressure solution, which is applicable in nature over long time scales and in chemically active environments [Bos and Spiers, 2000; Niemeijer et al., 2008]. These experiments follow work by Niemeijer et al. [2008], and show three lines of evidence for pressure solution: 1) linear stress relaxation with log time during holds to near zero shear stress, 2) high healing rates up to 0.15 $\Delta \mu$ per decade, and 3) porosity loss to < 10%, well below non-chemical compaction limits for granular materials. Our observations suggest that pressure solution starts anew when the fault zone is disrupted, as significant compaction occurs when $\sigma_n$ is increased (0.5 to 15 MPa; Figure 2-4) and following shear (Figure 2-2). Increased load likely alters the NaCl concentration gradient, prompting further erosion at grain contacts. Whereas shear likely disrupts the old framework of load-supporting grain necks, replacing them with new, sharp grain contacts that are prone to rapid pressure solution.

In general, porosity loss with pressure solution drives linear increases in the elastic wave speeds and amplitudes. Many previous works show that velocities generally increase linearly with $\phi$ loss [Hadley, 1976; Scott et al., 1993; Popp and Kern, 2001; Schubnel et al., 2003; Fortin et al., 2005; Fortin et al., 2007; Croizé et al., 2010], and figure 2-6 shows that $V_P$ and $V_S$ approach the mineral velocities at 0% $\phi$ [Nur, 1995; Popp and Kern, 1998; Mavko, 2009]. The relationship between $\phi$ and elastic wave speeds is driven by increased sample stiffness with lower $\phi$, thus increasing the bulk and shear moduli [Mavko, 2009]. Amplitude changes instead occur by density contrast changes between the steel forcing blocks and the sample layers, and are thus primarily dependent on sample porosity [Stein and Wysession, 2003].

The role of shear

Like compaction, $\phi$ is the predominant control on elastic wave speed and amplitude changes during fault slip (Figures 2-4, 2-6, 2-8, and 2-9). With early shear (0 to $\sim$ 2 mm) the fault layer dilates significantly (e.g., from 9 to 17% $\phi$; Figure 2-10). Dilation is a well-established granular response to shear [Mead, 1925; Marone et al., 1990]. The relationship of elastic wave speeds and amplitudes with $\phi$ during fault slip is best
observed following the short holds in p3636 (Figure 2-13). However, during the run-in or after long holds, this relationship breaks down following initial slip (Figures 2-13 and 2-14). In the following section we examine the potential effect of fabric formation on elastic wave speeds and amplitudes, and consider the utility of ultrasound as a probe for fault fabric development.

The run-in: from distributed to localized shear

During the 10 mm run-in the gouge-layer immediately dilates with shear (from ‘a’ to ‘b’, Figure 2-10). This $\phi$ increase drives dramatic elastic wave speed and amplitude decrease. However, after $\sim 2$ mm of slip these trends begin to break down – note the slight inflection at point ‘b’ in figure 2-10. These changes are coincident with peak stress and peak dilatancy rate (Figures 2-10, 2-13 and 2-14), and likely reflect a transition from localized to distributed deformation [Marone and Scholz, 1989; Marone et al., 1990].

Dilation requires work against normal stress, and thus greater shear stress. From the relation $\tau = \tau_f + \sigma'_n \frac{\delta\phi}{\delta\gamma}$ [Marone et al., 1990], where $\tau_f$ is the frictional resistance and $\sigma'_n$ is the effective normal stress, it is clear that dilatancy rate is directly tied to shear stress. Prior to peak $\frac{\delta\phi}{\delta\gamma}$, the rate of dilation increases with strain, and thus the energy required for shear increases with strain, driving distributed deformation. For example, a “localized” zone during this period has higher strain, incrementally requiring more energy. Alternatively, neighboring less-strained zones require less energy, and are thus preferred for shear, driving distributed deformation. Following peak $\frac{\delta\phi}{\delta\gamma}$, however, dilatancy rate decreases with strain, requiring less energy for shear and promoting localization. Previous works show that localized zones (i.e. Riedel shears) tend to initiate near peak $\frac{\delta\phi}{\delta\gamma}$ [Marone and Scholz, 1989; Marone et al., 1990; Haines et al., 2013]. Figure 2-16 is a thin section SEM micrograph from sample p3620, showing very well developed networks of Riedel shears with some boundary parallel Y-shears [e.g., Niemeijer et al., 2010]. We note that in situ porosity is much lower than the porosity following sample removal and drying, which results in fabric opening.
The role of fabric development

After peak $\delta\phi/\delta\gamma$, the elastic wave speeds and amplitudes diverge from linearity with compaction (Figure 2-13 and 2-14). Indeed, $V_P$ and $V_S$ transition to a positive relationship with $\phi$, in apparent contradiction to previous studies [e.g., Nur et al., 1995; Popp and Kern, 1998]. Alternatively, amplitudes become more sensitive to compaction following peak $\delta\phi/\delta\gamma$. While comminution and changes in mean stress also affect elastic wave speeds and amplitudes, we find that fabric generation following peak $\delta\phi/\delta\gamma$ provides the most viable explanation.

Within a fault, localized shear zones accommodate most strain once they begin to form [Logan et al., 1992; Haines et al., 2013]. The three dominant types of fabric are R1 (Riedel shear), R2 and Y shear zones, and they occur over a wide range of materials and conditions. Similar to our observations (e.g. Figure 2-16), Niemeijer et al. [2010] documented a predominance of Riedel shears at low shear strains with halite ($\sim 2.5 \gamma$). These tend to occur at high angles to $\sigma_n$, sub-parallel to the sound wave propagation direction in our experiments.

Riedel shears confound our ultrasonic measurements, which are sensitive to local heterogeneities, but these anomalous trends may provide useful information about Riedel shear formation. Our $V_P$ and $V_S$ observations represent the fastest P- and S-waves to propagate through the fault layers and forcing blocks. In a case with distributed deformation, all direct travel paths between the PZTs are equally fast. However, with localized deformation pathways become preferentially faster. Therefore the velocity observations likely represent velocity changes within local zones, whereas the $\phi$ measurements represent bulk changes. Thus the apparent relationship between bulk porosity and elastic wave speeds becomes skewed.

In contrast to elastic wave speeds, amplitudes become more sensitive to $\phi$ following peak $\delta\phi/\delta\gamma$ (Figures 2-13 and 2-14), but may also be understood in terms of fabric development. In the homogeneous case, P-waves arrive *en masse*, making an impulsive amplitude signal. However, with fast and slow pathways created by fault fabric, the P-waves arrive at slightly different times, blurring the signal. This *energy splitting* should similarly occur for S-waves. In this way, velocities and amplitudes are
not controlled by the bulk porosity, but rather the porosity of the fastest travel paths and energy splitting, respectively.

Near peak dilation, as opposed to peak $\delta \phi / \delta \gamma$, elastic wave speeds decrease and amplitudes increase, but maintain their anomalous trends with $\phi$ (Figures 2-13 and 2-14). These changes may be controlled by fabric evolution, as Riedel shears tend to rotate from high to low angles from $\sigma_n$ with shear [Logan et al., 1992; Haines et al., 2013]. For example figure 4-16 shows rotated Riedel shears, which occur sub-parallel to the shear direction. With Riedel rotation the fastest velocity travel paths become longer and must eventually cross these zones, driving velocity decrease. As the fastest waves become slower, their arrival becomes closer to the slowest waves, limiting energy splitting and driving amplitude increase. While fabric rotation explains the velocity increase followed by decrease, it does not offer an explanation for the apparent positive linear relationship that develops with $\phi$ (Figures 2-13 and 2-14), a trend that appears to hold through periods of bulk dilation and compaction. Whether this trend is coincidental or consequential remains an open question.

At the end of the run-in (e.g., 8-10 mm), velocities and amplitudes resume expected trends with porosity (Figures 2-13 and 2-14). With high strain, we find it likely that fabric rotation sub-perpendicular to $\sigma_n$ [Haines et al., 2013], and thus the wave propagation direction, drives this change (e.g. Figure 4-16; Haines et al., 2013). This geometry diminishes preferred travel paths for sound waves, which must travel across the heterogeneous zones. With all waves travelling through developed fabric, any changes to bulk porosity are experienced by all waves and expected porosity-velocity trends are resumed.

As described in the previous section, ultrasonic measurements are highly sensitive to fabric generation, and may prove useful to monitor these changes in future studies. Additionally, the apparent development of steady state fabric during these experiments gives credence to the use of run-ins, historically used to ensure experimental reproducibility [Marone, 1998].
**Velocity evolution during shear: The coseismic analog**

The relationship between porosity and the ultrasonic signals is strong over short holds (1-300s) and their shear periods (Figure 2-14). This is different from the run-in period, where velocity-porosity relationships break down with shear. This behavior indicates that the steady state fabric, established during the run-in, is still active through bulk layer dilation.

In contrast, following long holds the elastic wave speeds and amplitudes exhibit similar behavior to the run-in. With initial shear following long holds (e.g., 3,000 s), $V_P$, $V_S$, $A_P$, and $A_S$ decrease with dilation, followed by increasing velocities and stronger amplitude decreases with further dilation (Figures 2-13 and 2-14), similar to the run-in period. This behavior indicates a new period of fabric formation, perhaps necessitated by welding between grains resulting from pressure solution.

**Conclusions and implications for natural-fault observations**

These experiments provide detailed observations of fault zone damage during fault slip and subsequent fault healing. Our elastic wave speed observations show behaviors similar to previous works near natural fault zones, including coseismic velocity decrease followed by recovery with log time. However, we observe much larger signals overall. Previous works in nature generally focus on regional velocity changes, which likely result from dewatering, fracture, crack opening, and regional stress changes. To date, fault-zone guided waves provide the highest resolution observations around faults, ~200 m [e.g., Li et al., 2003; Li et al., 2006], but these are wide relative to the fault zone, which can be as narrow as 50 cm or less [Faulkner et al., 2010]. Our observations monitor the fault zone itself, and we observe very large elastic wave speed and amplitude changes. These changes are driven by dilation [Mead, 1925] and fabric formation during shear, and compaction via pressure solution during interseismic periods.

Porosity is the dominant control on elastic wave speeds and amplitudes throughout these experiments. Per 1% $\phi$ loss, $V_P$ generally increases by 3%, $V_S$ by 2%, $A_P$ by 10%, and $A_S$ by 7% (Figure 2-12), and these trends largely hold with shear driven dilation (Figure 2-6). Through the healing periods, $\phi$ generally decreases by ~25%/dec.
(Δ%), and occurs in tandem with strength increase (≈ 0.15 Δµ/dec.). This work indicates that changes to \( V_P \) and \( V_S \) through interseismic periods may be a useful tool to measure fault strength changes. Indeed, figure 2-15 shows that per 1% \( V_P \) and \( V_S \) change, strength might increase by ≈ 0.01 and ≈ 0.02 Δµ, respectively. Similar experiments could be used to catalog strength proxy observations for other fault-zone materials. To estimate fault strength changes in nature, however, high-resolution acoustic measurements must be made through a given fault zone, perhaps with borehole observatories similar to SAFOD [e.g., Niu et al., 2008].

These experiments may provide insight on velocity evolution during fault rupture, highlighting the potential importance of fault dilation and fabric formation for elastic wave speeds and amplitudes. These observations were limited to stable slip, and thus may not show some coseismic processes, however coseismic deformation is difficult to observe in detail due to the high slip velocities, small rise times, and noise from seismic energy. Our results suggest that changes to \( V_P \) and \( V_S \) do not scale with total fault slip, since most dilation occurs with early slip (≈ 0.4 γ in these experiments). This is clearly not the case for velocity decrease around natural faults, however, where velocity loss is more likely related to damage accumulation, perhaps with direct ties to earthquake magnitude [e.g., Brenguier et al., 2008; Li et al., 2003, Rubinstein et al., 2005; Nakata and Snieder, 2011].

Our results suggest that ultrasonic velocities and amplitudes may provide a proxy to monitor fabric formation within fault zones. While porosity changes are the predominant control on elastic wave speeds and amplitudes, fabric formation can dominate this signal, influencing expected velocity-porosity relationships. During fabric formation, the fault gouge develops heterogeneities. These localized zones cause sound waves to follow preferred travel paths, which results in velocity increase and amplitude decrease as these zones develop. Observations of subsequent velocity decrease and amplitude increase highlight Riedel rotation to near perpendicular with \( \sigma_n \). At high strains the elastic wave speeds and amplitudes resume expected trends with \( \phi \), indicating that steady state fabric has been developed. We observe fabric formation during the run-in and
after long holds, and similar processes likely occur in young faults and in faults with chemically assisted healing.
Figure 2-1. These double-direct shear experiments were conducted in a biaxial forcing rig and were brine saturated. This geometry is schematically shown in panel A. We collected continual active-source ultrasonic data ($V_P$, $V_S$, $A_P$, and $A_S$) throughout these experiments. An example seismogram is shown in panel B. We used S-wave piezoelectric transducers (0.5 MHz), which generate a small P-wave by mode conversion. The source piezoelectric transducer was mounted in the right side block and was excited by 900 V pulses. The receiver transducer was mounted in the left side block, with an excitation response generally less than 1 V. In general, the S-wave is much larger than the P-wave coda, and the P-wave is much larger than background noise. These features simplify P- and S-wave arrival picking.
Porosity (A) and seismogram (B) evolve dramatically under virgin compaction ($\sigma_n = 0.5$ and 15 MPa) and with applied shear load. Experiment p3635 was conducted on pure halite ($<125$ mm) in saturated NaCl brine. The sample was compacted for ~1 hour at 0.5 and for another hour at 15 MPa. The fault layers were then sheared 10mm (10 mm/s), held for 3,000s, and sheared again for 2mm (10 mm/s). Pressure solution is the dominant compaction mechanism in these experiments, and is responsible for the large drop in porosity from 50% to 10% over the course a few hours. The dramatic decrease in P- and S-wave flight time throughout these experiments is due to a combination of velocity increase and layer thinning, which was taken into account for velocity calculations.
Figure 2-3. Arrival times of P- and S-waves were picked semi-automatically using cross-correlation. For each experiment, we hand picked a master P-wave (A) and a master S-wave (C). These master waves were compared against a moving window for each of the other 1,000+ seismograms in the experiment (B and D, respectively). For each seismogram, the P- and S-wave was picked according to the highest correlation coefficient. Amplitudes ($A_P$ and $A_s$) were picked as the largest value following the P- and S-wave arrivals. Hand picking the S-wave arrival from the P-wave coda can be difficult if the waves are in phase. However, since the layer is continually thinning in these double-direct shear experiments, the S-wave migrates through the P-wave coda, and eventually arrives out of phase (D). This characteristic makes it easy to pick a master S-wave, which is used for the remainder of the picks.
Experiment p3455 was a simple compaction experiment conducted on saturated halite gouge, similar to the initial load up periods of the other experiments. The sample compacted with $\log(t)$ from 50% to 30% porosity within 1 hour at 0.5 MPa. Over this same period, the velocities and amplitudes increased dramatically. At 15 MPa, the sample experienced another wave and porosity decreased to 8% after three hours. Again, the velocities and amplitudes increased throughout this period. In general, these ultrasonic variables reflect changes in porosity, tending to increase linearly with porosity loss. Correlation coefficients for P- and S-waves are generally high (>0.97), but tend to be slightly weaker with smaller amplitudes (i.e. $A_P$).

**Figure 2-4.** Experiment p3455 was a simple compaction experiment conducted on saturated halite gouge, similar to the initial load up periods of the other experiments. The sample compacted with $\log(t)$ from 50% to 30% porosity within 1 hour at 0.5 MPa. Over this same period, the velocities and amplitudes increased dramatically. At 15 MPa, the sample experienced another wave and porosity decreased to 8% after three hours. Again, the velocities and amplitudes increased throughout this period. In general, these ultrasonic variables reflect changes in porosity, tending to increase linearly with porosity loss. Correlation coefficients for P- and S-waves are generally high (>0.97), but tend to be slightly weaker with smaller amplitudes (i.e. $A_P$).
Figure 2-5. Like the simple compaction experiment, porosity decreased and $V_P$, $V_S$, $A_P$, and $A_S$ increased during initial loading, shown here for experiment p3635. After this, the sample generally dilates during shear and compacts during holds. The velocities and amplitudes tend to inversely reflect the porosity evolution. In general, the absolute error for porosity and the velocities is large, owing to the absolute error for layer thickness ($\pm 0.1$ mm) and modeled sample mass. Since $A_P$ and $A_S$ are measured directly, their error is derived solely from electronic noise.
Figure 2-6. Over the course of a full experiment, $V_P$ and $V_S$ exhibit a linear relationship with $\phi$. This trend generally holds whether the sample is compacting, dilating, under shear, or being held. Our results, shown here for p3635, compare favorably with the pre-existing triaxial dataset from Popp and Kern [1998], conducted on natural rock salt samples.
Figure 2-7. Geometric thinning is a known drawback of many direct-shear experimental geometries, including biaxial and triaxial sawcut tests. With every increment of shear, a small amount of sample material is left behind. Following Scott et al. [2004], we modeled this loss assuming a rectangle and triangle of mass left behind (B). We validated this model against the actual amount of material lost for 14 experiments (A). In general, a rectangular model best predicts mass loss for these experiments. We used the rectangular model to estimate the sample mass and thus the sample porosity throughout each experiment.
Figure 2-8. This is a detailed plot of experiment p3635. With the first shear interval, the sample dilated dramatically, and the velocities and amplitudes decreased accordingly. During the 3,000 s hold the sample compacted and the elastic wave speeds and amplitudes increased. With subsequent shear the sample dilated, and the elastic wave speeds and amplitudes decreased. These variables generally scale with inverse porosity, but some second order effects occur during shear, and tend to coincide with peak stress and peak dilatancy rate (peak $\delta\phi/\delta\gamma$).
Figure 2-9. This is a detailed plot of experiment p3636, which contains multiple holds (1, 3, 10, 30, 100, 300, 3000 s), each followed by 2mm of shear. Changes in porosity, velocities, and amplitudes for virgin compaction and the first 10mm of shear were similar to p3635. These similarities largely extend to the second order changes in the elastic wave speeds and amplitudes during shear as well. Following initial and subsequent hold, the sample compacted, and the ultrasonic variables increased. During these shear periods, the sample dilated, and these variables tended to decrease. Additionally, hysteresis between shear and hold periods was limited, barring an overall porosity decrease over the course of the experiment.
Figure 2-10. This plot highlights changes in shear stress, porosity, and elastic wave speeds and amplitudes during the first shear period in p3635. Within the first 2 mm of shear (up until ‘a’), the sample dilated from 9% to 16% porosity and $V_p$, $V_s$, $A_p$, and $A_S$ decreased dramatically. Bulk ($K$) and shear ($G$) modulus are derived from $V_p$, $V_s$, and $\phi$, and tend to reflect changes in those variables. As expected, peak $\delta\phi/\delta\gamma$ coincides with peak stress [Marone et al., 1990]. However this point also marks a divergence in the trends of the elastic wave speeds and amplitudes. Porosity likely controls the velocities and amplitudes early with shear. Following peak $\delta\phi/\delta\gamma$ fabric development may play an important role for velocity and amplitude changes.
During holds, elastic wave speeds and amplitudes increase with log time, following $\phi$ loss. $A_p$ responds particularly strongly, increasing by 55%/dec. This figure highlights the 3,000s hold period of p3635. However, these log linear trend are not evident for the first 100-300 seconds, as adhesion is dominant over pressure solution at short hold times [Niemeijer et al., 2008].
Figure 2-12. Porosity, velocity and amplitude changes are consistent across our saturated halite experiments, even with varied stress and strain histories (table 1). In particular, this figure shows that the log linear changes in $\phi$, $V_p$, $V_s$, $A_p$, and $A_s$ is robust.
Figure 2-13. Elastic wave speeds and amplitudes scale linearly with inverse porosity, shown here for p3635. This is particularly true for both compaction periods: initial compaction (black lines, left panels) and the 3,000s hold (black lines, right panels). With shear following these compaction phases (gray lines), $V_p$ and $V_S$ generally decrease with dilation. However, following peak $\delta\phi/\delta\gamma$, these trends break down, as $V_p$ and $V_S$ gradually change to a positive linear relationship with $\phi$. Alternatively, elastic wave amplitudes become more sensitive to $\phi$ change over this period. Following peak $\delta\phi/\delta\gamma$, fabric evolution may modulate the velocities and amplitudes via wave splitting. Near the end initial shear period (green dots, left panels), elastic wave speeds and amplitudes appear to revert to their original trends with $\phi$. 
Figure 2-14. Like p3635, velocities and amplitudes trend with inverse porosity, shown here for experiment p3636. These general trends are present over multiple slide-hold-slides (1 to 3,000 s), and show remarkable hysteresis with compaction (holds) and dilation (shears). Like p3635, with shear following initial compaction and following the 3,000s hold, these trends break down, and seemingly revert to their original trends at the end of the 10 mm shear period. If these aberrant trends are related to fabric generation and evolution, it is possible that “steady state” fabric developed during the 10 mm of shear remains active after the short holds. Conversely, after long holds (i.e. 3,000s) the fabric may be disrupted significantly, and must be rebuilt with later shear.
Figure 2-15. During holds, fault strength (A) and elastic wave speeds (B) increase with log(τ). Elastic wave speeds directly reflect and depend on porosity change. Similarly strength increase depends on φ change, since work against normal stress is required as the sample dilates with shear. Therefore elastic wave speed changes might serve as a probe to measure in situ fault strength changes. In these experiments, strength increases by ~ 0.01 and ~0.02 Δµ per percent increase of $V_P$ and $V_S$. While these relationships are material and environment dependent, a catalog of these changes may be useful to estimate strength changes in natural faults with further development of velocity changes within fault zones.
Figure 2-16. Thin section SEM micrograph from experiment p3620 ($\gamma = 4.6$). The fault zone developed pervasive and through-going Riedel shear networks typical of granular materials [e.g. Logan et al., 1992; Niemeijer et al., 2010; Haines et al., 2013]. Riedel shears tend to rotate with shear strain [Haines et al., 2013] and, in some cases, become sub-parallel to the shear direction (perpendicular to P and S-wave propagation). Boundary parallel Y-shears are also evident and occur at the interface with the steel forcing block teeth.
### Table 2-1. Experiment details

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<th>$\sigma_n$ following run-in (MPa)</th>
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<th>Shear strain ($\gamma$)</th>
<th>Initial mass (g)</th>
<th>Final mass (g)</th>
<th>Measured mass loss (g)</th>
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Chapter 3: Deformation band formation and strength evolution in unlithified sand: The role of grain breakage

Abstract

We report on laboratory experiments designed to investigate the strength evolution and formation mechanisms of cataclastic deformation bands hosted in unlithified sand, with particular focus on the role of grain breakage. Cataclastic deformation bands are characterized by particle size reduction and increased resistance to weathering compared to parent material. We recovered bands intact from late Quaternary, nearshore marine sand in the footwall of the active McKinleyville thrust fault, Humboldt County, CA. Tabular samples 3 to 5 mm thick and 5 cm x 5 cm in area were sheared at normal stresses representative of in situ conditions, 0.5-1.8 MPa, sliding velocities from 10 μm/s to 10 mm/s, and to shear strain up to 20. Cataclastic deformation bands are stronger than parent material (coefficient of internal friction $\mu_i = 0.623$ and $\mu_i = 0.525$, respectively), and exhibit a peak strength followed by weakening. Parent material exhibits significant strain hardening; the frictional yield strength increases up to 9% for a shear strain of 10. Detailed particle size analyses show that strain hardening in parent material is coincident with increased fine particle abundance, resulting from pervasive grain breakage. Our results support the hypothesis that cataclastic deformation bands are stronger than their surrounding parent material due to shear-driven grain breakage during band formation. We suggest that the combination of strain localization during band formation and strain hardening on individual bands results in dense networks of deformation bands.

Introduction

In response to shear strain, porous granular materials often fail in tabular zones referred to as deformation bands. We study deformation bands that exhibit shear offset and are characterized by grain size reduction and increased resistance to weathering relative to the unlithified sand in which they occur. These bands share characteristics of deformation bands described in sandstone [Fossen et al., 2007] and thus we refer to them
as cataclastic deformation bands, or alternatively cataclastic shear bands. Previous researchers have argued that strain hardening inhibits additional shear on such bands once they form, resulting in networks of closely spaced bands [Aydin, 1978; Mair et al., 2000; Schulz and Balasko, 2003, Fossen et al., 2007]. This hypothesis requires that cataclastic shear bands are stronger than their parent material, and that parent material strengthens in response to shear driven cataclasis. While these components are generally consistent with existing data [e.g., Mair et al., 2000], the hypothesis has not been adequately tested in laboratory experiments. The purpose of our study is to present detailed laboratory measurements necessary to test this hypothesis, and to combine measurements of frictional properties with particle size analysis to investigate the processes responsible for cataclastic shear band formation and strain hardening in poorly lithified sands.

Cataclastic shear bands are best known in sandstone where estimated formation depths range from 1.5 to 2.5 km, at 20-40 MPa effective lithostatic stress [Antonellini et al., 1994; Fossen et al., 2007]. These deformation bands localize along zones of parent rock that has been weakened [Mair et al., 2000; Rawling and Goodwin, 2003], and deformation is often accommodated via cataclasis, with pervasive grain breakage. Cataclastic shear bands are commonly millimeters to centimeters thick, and accommodate less than a few cm of shear displacement. Cataclastic deformation band formation has also been reported during laboratory studies conducted on sandstone at 34 MPa confining pressure [Mair et al., 2000]. However, cataclastic deformation bands also occur at much shallower depths and confining pressures in unlithified materials [Cashman and Cashman, 2000; Wen and Aydin, 2004].

Mair et al. [2000] presented a conceptual model for cataclastic deformation band formation in field and laboratory settings. They address the problem of localization in a strain hardening material; that is, how can strain localize in a zone that is adjacent to weaker material? In particular, they argue that the parent material must first weaken in localized zones in order for deformation bands to form, but that strain hardening processes terminate shear within such zones. In sandstones, and in unlithified sands such as we studied, cataclasis seems to be an important process in both cataclastic shear band formation and shear termination, resulting in networks of closely spaced deformation
bands. Previous workers have suggested that stress concentrations between bands promote the formation of stepover cataclastic deformation bands, resulting in Riedel, ladder, and “radiator rock” arrays observed in the field [Schulz and Balasko, 2003].

Recent theoretical work has been conducted to identify the formation conditions of different types of deformation bands [Wong et al., 1992; Zhu and Wong, 1997; Borja, 2004; Borja and Aydin, 2004; Wong et al., 2004; Aydin et al., 2006]. A cap plasticity model is preferred to describe these formation conditions [Fossen et al., 2007]. When the material crosses the yield surface, or cap, a bifurcation occurs, and a deformation band will begin to form. The type of deformation band that forms is dependent on the position in stress space at the time of failure [Wong et al., 1992; Zhu and Wong, 1997; Borja, 2004; Borja and Aydin, 2004; Wong et al., 2004; Aydin et al., 2006].

Field site

We collected cataclastic shear bands from an unlithified, late Pleistocene, nearshore-marine sand unit [Cashman and Cashman, 2000] in Humboldt County, CA (Fig. 3-1). This site is a ~150 m wide zone (Fig. 3-2) that contains two varieties of deformation bands, dilation bands and cataclastic shear bands [Cashman and Cashman, 2000; Du Bernard et al., 2002]. The parent material is well-sorted sand that contains abundant feldspar grains and lithic fragments. These deformation bands formed in response to slip on proximal strands of the seismically active McKinleyville thrust fault [Clarke and Carver, 1992; Cashman and Cashman, 2000].

Maximum burial depth for the deformation bands we studied is less than 100 m, which corresponds to < 1.8 MPa effective lithostatic stress. Burial depth is inferred from the stratigraphic context of the parent material unit: it is bounded by unconformities, a molar of the mastodon Mammut Americana was found in the lowermost horizon of the unit (G.A. Carver, personal communication, 1993, cited by Harvey [1994]), a thermoluminescence age estimation of 176±33 ka was determined for a mud horizon near the top of the unit [Berger, 1992], and the unconformably overlying 83 ka marine terrace [Carver and Burke, 1992] is currently ~33 m above sea level. Together, these features record the presence of the parent material unit at or above paleo-sea level multiple times.
in the late Pleistocene; there is no evidence of significant burial.

Cataclastic shear bands occur in two conjugate sets (Fig. 3-2), dipping at approximately 30° to the north and south, respectively [Cashman and Cashman, 2000; Du Bernard et al., 2002]. The majority of these bands are < 1 cm thick and record reverse-fault shear displacement of a few mm to a few cm, corresponding to shear strains ($\gamma$) < 10 (Table 3-1). In a few cases, the bands are 10 to 20 cm thick and host a few meters of shear ($\gamma \approx 50$). Cataclastic shear bands at our study site weather in positive relief (Fig. 3-2), indicating greater resistance and strength relative to their parent material. At our field site, dilation bands are generally 1 to 2mm thick, subhorizontal, and are spatially associated with cataclastic shear bands. Dilation bands are zones of increased granular porosity, which have become infilled with clay [Du Bernard et al., 2002].

Relative to their parent material, our deformation bands are composed of finer and more angular grains, and have decreased porosity (Fig. 3-3 and 3-4; Table 3-1) [Cashman and Cashman, 2000]. Similarly, our experiments on parent material document cataclastic deformation with attendant compaction, as described below. These are identifiable characteristics of cataclastic shear bands formed in sandstone, hence we use that term here. Unlike Aydin et al. [2006], we argue that these bands have undergone significant grain breakage to accommodate shear. Grain breakage results in a wide particle-size distribution (PSD) relative to the parent material (Fig. 3-3), and is consistent with the observations of Rawling and Goodwin [2003].

**Strength of cataclastic shear bands**

We measured the frictional shear strength of cataclastic shear bands and their parent material. In addition, we analyzed the evolution of parent material strength and PSD as a function of shear strain. We specifically assessed the effect of grain breakage on parent material strength. Shear-driven grain breakage may enhance material strength as a result of increased grain angularity [e.g., Zhu et al., 1997; Cresswell and Barton, 2002; Mair et al., 2002; Anthony and Marone, 2005; Guo and Morgan, 2006], pore collapse [Knudsen, 1959; Aydin, 1978], and changes in PSD [Lambe and Whitman, 1969]. Cementation may also play an important role in deformation band evolution.
Permeability and fluid flow

Deformation bands have received significant attention due to their influence on fluid flow and reservoir systems [Zhu and Wong, 1997; Aydin, 2000; Lothe et al., 2002; Parnell et al., 2004; Perez et al., 2010]. The porosity of cataclastic shear bands relative to parent material can be reduced by an order of magnitude or more. The permeability of these bands may be anisotropic and significantly lower than parent material. [Antonellini et al., 1994; Zhu and Wong, 1997; Lothe et al., 2002; Fossen et al., 2007]. Within a cataclastic shear band, permeability reduction is a result of PSD changes, tighter packing, inter-grain crack closure, increased tortuosity, and, in some cases, pore collapse immediately outside the band. Laboratory measurements of permeability perpendicular to our cataclastic shear bands are 1 to 3 orders of magnitude lower than that of parent material [Perez et al., 2010]. At our field site the preferred orientation of colored streaks, authigenic iron oxide deposits, is altered near the deformation bands (Fig. 3-2), indicative of their effect on fluid flow.

Methods

We identified cataclastic shear bands for collection along an actively eroding sea cliff that exposes the un lithified, late Pleistocene parent material unit (Fig. 3-1). To expose unweathered bands for collection, the outcrop surface was first removed using hand tools. Individual bands were extracted and sealed in plastic wrap to preserve sample moisture and integrity. Our suite of samples represent cataclastic deformation bands over a range of shear displacements, orientations, and thicknesses (Table 3-1).

Sample characterization

At the field site we measured thickness and apparent displacement of the cataclastic shear bands. Shear displacement was constrained by apparent lateral offset of sedimentary strata and other bands. Shear strain was estimated by dividing the apparent offset by the band thickness (Table 3-1).

Figure 3-3 shows particle size distributions for bands and parent material. To
highlight differences between samples we calculated inclusive graphic skewness, $\lambda$ [after Folk and Ward, 1957], using the mean size of the parent material (205 $\mu$m) in lieu of the mean size of the material of interest. In this scheme, a PSD skewed toward fine particles has a greater skewness.

Porosity ($\phi$) of intact bands and parent material was calculated from bulk density and solid density, following $\phi = 1 - (\rho_b / \rho_s)$ (Table 3-1). The solid density, $\rho_s$, was measured by weighing a sample and then measuring its displacement in water, allowing the water to permeate the pore space. The bulk density, $\rho_b$, was measured by weighing a sample, applying a waterproof acrylic coating, then measuring its displacement in water, preventing the water from permeating the pore space. We assume that the density difference between $\rho_s$ and $\rho_b$ is due to pore space.

**Laboratory sample preparation**

Layers 3 to 5-mm thick and 5 cm x 5 cm in area were extracted from the field samples for laboratory experiments. For most bands, we were able to recover multiple layers, which were used to assess experiment reproducibility and in parametric studies of strength as a function of experimental variables such as normal stress (Table 3-2). Layers were first rough cut and then trimmed to specific layer dimensions with a hand-held, precision rotary sander. Layers of loose parent material were constructed to specific layer dimensions in a leveling jig. Layers were oven dried prior to shear (100° C, >30 minutes) for dry experiments or they were presaturated with tap water for saturated experiments.

**Laboratory techniques**

We conducted friction experiments in a servo-controlled biaxial testing apparatus [for details see Anthony and Marone, 2005; Savage and Marone, 2007; or Rathbun and Marone, 2008]. Sample layers where sheared in a three forcing-block configuration using either double-direct shear, DDS, or single-direct shear, SDS (Fig. 3-5). For DDS experiments two layers were sheared simultaneously. For SDS experiments a frictionless bearing was used in place of one layer [e.g., Marone, 1995; Marone et al., 2008]. Normal stress on the layers was maintained constant via a fast acting servo-hydraulic controller.
Layers were sheared by driving the central block of the sample assembly down at a controlled rate, typically 10 µm/s. Shear and normal stresses were measured with strain gauge load cells accurate to ±0.1 kPa, and displacements were measured with Direct Current Displacement Transducers (DCDT's) accurate to ±0.1 µm. Data were recorded digitally at 10 kHz with a 24-bit system and averaged to 10 to 100 Hz for storage.

We compared the strength of cataclastic deformation bands and parent material using the SDS configuration (Fig. 3-5a and Table 3-2). This configuration required minimal sample material, and was used to maximize data from a limited number of samples.

Experiments conducted with the DDS configuration were designed to assess the effect of grain breakage and changes in PSD on parent material strength. This suite of experiments was conducted on parent material at constant normal stress ($\sigma_n = 0.75-1.75$ MPa), moisture conditions (dry, saturated) and driving velocity (10-10,000 µm/s) to variable final shear strains ($\gamma = 0.5-20$; Table 3-3). We measured PSD at the beginning and end of all experiments. Initial layer thickness was measured with calipers to ±50 µm. Change in layer thickness was measured during each experiment, and was corrected for geometric thinning [e.g., Scott et al., 1994]. We used layer thickness change to assess PSD changes during an experiment, assuming that poorly sorted PSDs can compact more than well-sorted distributions. Details of the layer thickness change as a function of shear indicate a combination of shear driven dilation and compaction [e.g., Mair et al., 2002; Anthony and Marone, 2005; Rathbun and Marone, 2008].
All experiments began with a shear displacement of 5 mm at a normal stress of 0.5 MPa. This 'run-in' was designed to minimize differences in initial packing state, and to develop a consistent initial fabric [e.g., Marone, 1998]. Our data include measurements of the bulk shear strain, derived by integrating the incremental shear displacement divided by the instantaneous layer thickness. At the end of all experiments, shear stress and normal stress were removed, and the deformed layers were collected. Residual sample wafers from the saturated experiments were sectioned for SEM analysis. Sample collection was not possible for dry experiment because the layers lacked cohesion.

**Experiment results**

Cataclastic shear bands have distinctly different frictional histories than parent material (Fig. 3-6). Upon shear loading, all materials exhibit an initial increase in stress associated with elastic loading of the layer, followed by permanent inelastic shear deformation, consistent with previous observations [e.g., Feda, 1982; Bonn and Denn, 2009]. Cataclastic shear bands typically yield at higher shear stress than parent material. Following yield, parent material often exhibits strain hardening. In contrast, cataclastic shear bands exhibit a peak stress followed by weakening. In general, steady state strength of the bands was 90-97% of the peak stress and was attained after a shear strain of ~1-2.

**Comparison of cataclastic shear bands and parent material**

Cataclastic shear bands are stronger than parent material under all of the conditions we studied (Fig. 3-7). Coulomb-Mohr failure envelopes were determined by fitting our measurements with the relation $\tau_f = \mu_i \sigma_n + C_0$, where $\mu_i$ is the coefficient of internal friction, $\sigma_n$ is normal stress on our layers, $C_0$ is cohesion, and $\tau_f$ is the maximum shear stress at the initiation of permanent shear deformation. Herein $\tau_f$ is defined as the maximum shear stress over the interval from the yield point to an additional shear strain of 1. Figure 3-7 shows that cataclastic shear bands ($\mu_i = 0.623, C_0 = 22$ kPa) have greater internal friction and similar cohesion compared to parent material ($\mu_i = 0.525, C_0 = 48$ kPa); see also Table 3-2. Field evidence indicates that cataclastic shear bands have
greater cohesive strength than parent material, given their increased resistance to weathering (Fig. 3-2); however the magnitude of this difference is apparently within our experimental measurement uncertainty.

The strength of parent material generally increases as a function of shear strain (Fig. 3-6), and such strain hardening is observed under all experimental conditions (e.g., normal stress, sliding velocity, dry/saturated). The degree of strain hardening is variable, with sliding friction increasing by 0 to 1% per unit increase in shear strain, but it is consistently positive when averaged over shear strain greater than ~0.2. Over a full experiment the strength of parent material approaches, but seldom reaches, cataclastic shear band strength (Fig. 3-8). Figure 3-8 compares the strength of the cataclastic shear bands to the strength of parent material for the duration of each SDS experiment ($\gamma = 0$-20). The DDS experiments were designed to explore strain hardening of parent materials in greater detail.

**Strain hardening behavior of parent material**

Parent material exhibits systematic strengthening and grain breakage as a function of shear strain under all conditions (Fig. 3-9 and 3-10). The particle size of undeformed parent material is primarily 190-220 µm. Figure 3-9 shows decreased abundance of coarser grains (190-220 µm) and increased abundance of finer grains (1-100 µm) as a function of shear strain. Figure 3-10 shows that parent material skewness and strength increase as a function of strain over a range of normal stresses (0.75-1.5 MPa). Strength increase and fine particle generation occur most readily early in experiments ($\gamma = 0$-10) and become less pronounced as the experiments progress (Fig. 3-10). In general, strengthening and fine particle generation are enhanced at higher normal stresses. Parent material strength and skewness (more fine particles) are positively related (Fig. 3-10, insets).

Fine particle generation is dramatically enhanced under saturated conditions. Figure 3-11 shows that saturated experiments have a pronounced abundance of fine grains (1-100 µm) relative to dry experiments, under otherwise identical experimental conditions. The PSDs from saturated experiments are similar to the cataclastic shear
bands, both in terms of the 190-220 µm and 1-100 µm size fractions (Fig. 3-11). Sheared parent material also appears similar to cataclastic deformation bands in thin section (Fig. 3-12); the large grains tend to be subrounded, whereas the smaller grains tend to be angular. As well, sheared parent material and cataclastic shear bands have grain supported skeletons, where many grain contacts are surrounded by fine-grained matrix. The PSD in figure 3-11 is representative of cataclastic shear bands and shows a distinct 0.3-0.4 µm size fraction. This size fraction is not observed in sheared parent material under any experimental conditions (Fig. 3-11), which is a point we discuss below.

Sliding velocity does not have a systematic effect on the behavior of parent material for our range of experimental conditions (Table 3-3). In particular, we analyzed the effect of velocity on $\tau_f$ and the rate of strain hardening, and did not find an effect. Inherent variation between samples, due to material heterogeneity or experiment procedures, may have a greater effect on parent material behavior than slip velocity.

Discussion

Our results show that cataclastic shear bands are stronger than parent material. This strength difference is particularly evident at the onset of permanent shear deformation (Fig. 3-6 and 3-7). These data imply that, once cataclastic shear bands form, they will not reactivate to accommodate additional strain. Instead, additional strain will be accommodated within nearby parent material. An intuitive result of this behavior is networks of closely spaced cataclastic deformation bands, as has been posited by previous researchers for sandstones [Aydin, 1978; Schulz and Balasko, 2003]. Our results demonstrate that this explanation also applies to unlihified sand.

The role of grain breakage

Grain breakage of parent material occurs over our full range of experimental conditions as indicated by the decreased abundance of coarser grains (190-220 µm) and increased abundance of finer grains (1-100 µm) with strain. Grain breakage is observable during parent material experiments in the form of short-term chaotic frictional behavior (Fig. 3-6). Similar data have been described previously [Feda, 2002; Lobo-Guerrero and
Vallejo, 2005; Guo and Morgan, 2006]. Feda [2002] describes these as garland like friction curves, which are attributed to grain breakage or cyclic softening associated with force chain failure and reorganization of a new load bearing skeleton.

Progressive grain breakage results in strain hardening of parent material as indicated by the positive correlation of skewness (more fine particles) and strength (Fig. 3-10). Grain breakage results in strain hardening in three possible ways: 1) It generates finer grains (Figs. 3-9 and 3-11; Table 3-3), which are expected to be stronger than larger grains, according to Petch's law [Petch, 1953]. 2) Finer grains fill in pore space, effectively increasing the grain coordination number and widening the PSD, each of which cause strengthening [Sammis et al., 1987]. 3) Grain breakage generates angular grains, which have been found to increase the frictional strength of granular materials [Lambe and Whitman, 1969; Mair et al., 2002; Anthony and Marone, 2005]. In our experiments, angular grains are generated by fracture of the subrounded parent grains (Fig. 3-12).

Grain breakage readily occurs early in experiments and becomes less prominent with strain. This is indicated by the large initial ($\gamma = 0-5$) increase in skewness followed by much less change at higher strains (Fig. 3-10). Deformation via grain breakage decreases in favor of frictional grain boundary sliding, following the least energy intensive deformation mechanism. Grain breakage makes frictional sliding less energy intensive, because dilation of finer grains, if shear is localized, requires a smaller increase in layer thickness. In contrast, further grain breakage becomes more energy intensive, because finer grains are generally stronger, and the grain coordination number is increased. In effect, the fine grains generated by grain breakage promote dilation and frictional grain boundary sliding instead of continued grain breakage.

Cataclastic shear band formation

Parent material begins to develop a number of deformation-band-like characteristics during our experiments. In particular, parent material approaches cataclastic deformation band strength, especially at high shear strains ($\gamma = 20$; Fig. 3-8). The particle size distribution of sheared parent material becomes similar to that of
cataclastic deformation bands (Fig. 3-11). In addition, sheared parent material becomes visually similar to the bands, including many large grains surrounded by pore space that is filled in by matrix (Fig. 3-12). In particular, both sheared parent material and cataclastic shear bands have grain supported skeletons, where many grain contacts are surrounded by fine-grained matrix.

Under dry experimental conditions the frictional strength of parent material approaches, but does not reach of the strength of cataclastic shear bands (Fig. 3-8). The same is true for the particle size distribution of parent material and the cataclastic shear bands (Fig. 3-11). The differences in strength and PSD are likely due to differences in laboratory versus natural sliding conditions (e.g., diagenesis, cementation, or sliding velocity). One could argue that shear strain is a factor; however our laboratory experiments reached shear strains of up to 20, which is similar to the field estimates for the natural bands. Normal stress is another potential variable, however our experiments were conducted at normal stresses near the upper end of the likely range of in situ normal stress (1.8 MPa), and thus this does not appear to be a likely explanation.

Under saturated experimental conditions parent material develops a PSD remarkably similar to cataclastic shear bands (Fig. 3-11). This similarity is highlighted by a decrease in the 190-220 μm and increase in the 1-100 μm size fraction relative to dry experiments under otherwise identical experimental conditions. These PSD differences arise from an increased propensity for granular fracture under saturated conditions, which is consistent with the effect of hydrolytic weakening on fracture [e.g., Miura and Yamanouchi, 1975].

Cataclastic shear bands have a greater proportion of fine particles, in the range 0.3-0.5 μm, than virgin or sheared parent material (Fig. 3-11). These particles represent a fraction below the 1 μm mechanical grinding limit suggested by Sammis and Ben-Zion [2008]. Diagenesis and/or clay accumulation by filtration of ground water could account for fine particles, and would indicate that cataclastic shear bands evolve through time. Subcritical crack growth and fragmentation, aided by hydrolytic weakening, could also account for particles finer than 1 μm, and might occur shortly after band formation.
Another possible explanation for particles finer than 1 μm is coseismic, shock loading; however we do not have additional supporting evidence for this.

The role of diagenesis

We posit that cementation accounts for the differences in strength and PSD between the cataclastic shear bands and experimentally sheared parent material. Our data show that the bands have higher initial frictional strength than parent material (Fig. 3-6). Cataclastic shear bands also exhibit post-peak weakening, in contrast to the strain hardening observed in parent material. Cementation may arise in nature from iron oxide and clay accumulation [Cresswell and Barton, 2002], which may be enhanced in the bands due to their decreased porosity and permeability relative to parent material. Du Bernard et al. [2002] have argued that clay accumulation plays a role in dilation bands, and thus it may be a factor in our shear bands. Similarly, iron oxide precipitation is pervasive at our study site (Fig. 3-2) and may be enhanced in cataclastic deformation bands via preferred water retention, which may increase the kinetics of chemical weathering and enhance cementation.

Strain hardening and localization within cataclastic shear bands

A fundamental question about the mechanics of cataclastic shear bands involves how localization can be sustained in a strain-hardened zone that is adjacent to weaker material. For cataclastic shear band formation in sandstones, previous workers have noted the importance of an initial strain weakening mechanism such as cohesive breakdown [Mair et al., 2000; Lothe et al., 2002]. However, the application of cohesive breakdown to un lithified sand is unclear. In particular, our experiments do not show consistent strain weakening behavior of parent material. In their present state, our deformation bands are stronger and have enhanced cohesive strength compared to the surrounding parent material. However, they presumably formed from parent material, prior to acquiring increased cohesion. Our experiments suggest that cataclastic shear band formation is associated with either grain fracture, causing local weakening, or a mechanism associated
with the propagation of a coseismic shear rupture from a nearby lithified unit [e.g., Cashman and Cashman, 2000; Wen and Aydin, 2004; Cashman et al. 2007].

Conclusions

Our results are consistent with the hypothesis that cataclastic deformation bands are stronger than their parent material. We find that the relative strength of cataclastic shear bands is largely a result of strain hardening via shear driven grain breakage. Pervasive grain breakage results in strengthening via increased grain angularity and fine particle abundance. As shear progresses, grain breakage (and strength increase) abates in favor of steady-state grain boundary sliding. Although grain breakage results in increased material strength, it cannot account for the full strength difference between cataclastic shear bands and their parent material. Given this strength discrepancy and the presence of clay sized particles (< 1 µm), we posit that cementation acts to further increase the strength of cataclastic deformation bands (Fig. 3-13). Preferential cementation of shear bands, relative to the surrounding parent material, has important implications for flow paths, permeability, and strength in poorly lithified sands.
Figure 3-1. Map of site and fault surface traces near the McKinleyville area, California. Modified from Cashman and Cashman [2000].

Figure 3-2. Photograph of deformation bands expressed in relief at our field site. Deformation bands occur within unlithified, late Quaternary, nearshore marine sand in the McKinleyville Fault footwall (see Fig. 3-1). Color streaks (subvertical) on the outcrop are due to the deposition of authigenic iron oxide minerals in pore space. Outcrop faces west; north is to the left in this photograph. Note pencil for scale.
Figure 3-3. PSDs of parent material and cataclastic shear bands. Shear strain is indicated as $\gamma$ (unknown for sample C06). PSDs were determined using a Malvern Mastersizer S (dynamic particle size range from 0.05–900 $\mu$m). See Table 1-1 for other details of sample characteristics.
Figure 3-4. Surface SEM images of (A) intact shear band 08B03 (γ = 45) and (B) intact parent material. Scale bars = 500 µm.

Figure 3-5. Experimental forcing block geometries for (A) single-direct shear (SDS) and (B) double-direct shear (DDS)
Figure 3-6. Coefficient of sliding friction as a function of shear strain for representative experiments conducted on parent material (black lines) and cataclastic shear bands (gray line). For each experiment, we show normal stress (MPa) and experiment number (see Table 2 for other details). Maximum shear stress at the initiation of shear ($\tau_f$) is shown for experiment p1982.

Figure 3-7. Coulomb-Mohr failure envelope showing $\tau_f$ as a function of normal stress for cataclastic shear bands ($\mu_i = 0.623$, $c_0 = 22$ kPa) and parent material ($\mu_i = 0.525$, $c_0 = 48$ kPa).
Figure 3-8. Range of friction values for parent material and cataclastic shear bands. Parent material friction values are shown for the initial portion of individual experiments (black bars) and for the remainder of the experiment (gray bars). Black dots show the maximum friction at the initiation of shear for cataclastic shear bands ($\tau_f$ divided by normal stress).

Figure 3-9. PSDs of parent material as a function of shear strain from dry experiments at $\sigma_n = 0.75$ MPa and shear velocity of 10 $\mu$m/s; 220 $\mu$m represents the most abundant size. Note that the peak abundance decreases systematically as a function of shear strain. This decrease abates with increasing shear strain.
Figure 3-10. [Main plots] Friction (circles) and skewness (triangles) as a function of shear strain at normal stresses of (A) 0.75, (B) 1.25, and (C) 1.5 MPa. Skewness and friction values are reported from the end of individual experiments ($\gamma = 0-20$). Details of skewness and friction as a function of $\gamma$ are shown at each normal stress. [Insets] Friction as a function of skewness at the end of each experiment (squares). Details of friction as a function of skewness are shown at each normal stress.
Figure 3-11. Particle size distributions of parent material, a cataclastic shear band (08B04; $\gamma = 5$), and parent material from saturated (p2130; $\gamma = 6-7$) and dry (p2141; $\gamma = 7.65$) experiments conducted at 0.75 MPa normal stress. The PSD from the saturated experiment shows fewer coarse grains and an increased abundance of fine grains (1-100 $\mu$m) relative to the dry sample (p2141). The PSDs for the saturated experiment and cataclastic shear band are similar.
Figure 3-12. Oriented SEM images of (A) shear band 08C05 ($\gamma = 6$) and (B) parent material from saturated parent material experiment p2130 ($\gamma = 6$-$7$; $\sigma_n = 0.75$ MPa). Both materials show grain shape similarities. Arrows denote sense of shear. Scale bars = 500 µm.

Figure 3-13. Possible evolution paths ($\mu$, skewness and permeability) of cataclastic shear bands developing from parent material. Model A is a case where band characteristics develop due to grain breakage. Model C is a case where band characteristics develop due to cementation. Model B, our preferred model, is a case where cataclastic shear band characteristics develop initially by grain breakage, and are later modified by cementation.
Table 3-1. Physical properties of shear bands and parent material collected from the study site.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Offset (mm)</th>
<th>Thickness (mm)</th>
<th>Shear Strain</th>
<th>Porosity (%)</th>
<th>Mean Grainsize (µm)</th>
<th>Skewness</th>
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</thead>
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<td>na</td>
<td>na</td>
<td>44.47</td>
<td>200</td>
<td>0.29</td>
</tr>
<tr>
<td>08C05</td>
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<td>12.5</td>
<td>6</td>
<td>25</td>
<td>35</td>
<td>0.98</td>
</tr>
<tr>
<td>C06</td>
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<td>unkn</td>
<td>27.5</td>
<td>103</td>
<td>0.90</td>
</tr>
<tr>
<td>08B04</td>
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<td>9.0</td>
<td>8</td>
<td>30.5</td>
<td>45</td>
<td>0.95</td>
</tr>
<tr>
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<td>50</td>
<td>24.5</td>
<td>50</td>
<td>0.94</td>
</tr>
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</table>
Table 3-2. Single-direct shear experimental parameters. Experiments were conducted on dry parent material and cataclastic shear bands at constant sliding velocity (10 µm/s) and normal stress (0.5-1.85 MPa). We report the maximum shear stress ($\tau_f$) and friction ($\mu_f$) at the initiation of permanent shear deformation.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Material</th>
<th>$\sigma_n$ MPa</th>
<th>$\tau_f$ MPa</th>
<th>$\mu_f$</th>
</tr>
</thead>
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</tr>
<tr>
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<td>Parent material</td>
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<td>0.614</td>
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<td>0.571</td>
</tr>
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<td>0.572</td>
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Table 3-3. Double-direct shear experimental parameters. Experiments were conducted on parent material at constant sliding velocity (10 μm/s), normal stress (0.75-1.75 MPa), and moisture conditions (dry/saturated). We report the final shear strain, friction, and skewness for each experiment.

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<tr>
<th>Experiment</th>
<th>$\sigma_{ns}$ (MPa)</th>
<th>Shear strain ($\gamma$)</th>
<th>Sliding velocity, $\mu$m/s</th>
<th>Saturated</th>
<th>Skewness</th>
<th>$\mu$ at final strain</th>
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Chapter 4: Permeability and mechanical properties of halite-clay-quartz bearing faults with compaction and shear

Abstract

Faults in marine-sediment basins often dictate fluid flow and act as petroleum traps or seals. Permeability contrasts between these zones and the surrounding country rock can be large, depending on variables like fault composition, stress, and strain history. Despite the importance of such faults, our understanding of their frictional properties and permeability is relatively poor. In particular, the role of halite and clay within faults adjacent to salt domes and the possibility for multimechanism behavior, including brittle deformation at high strain rates and ductile deformation and pressure solution at slower rates, is unclear. Here, we report on a suite of laboratory experiments to assess the roles of fault composition, load, and shear strain for fault permeability and frictional properties. We studied synthetic fault gouge composed of mixtures of quartz, halite, and clay composed of illite shale or smectite (montmorillonite), with maximum end-members for each material up to 100% quartz, 54% clay, or 90% halite by mass. We measured fault permeability under compaction at 2, 4, and 6 MPa effective normal stress, $\sigma_n$', and at low (2-5) and high (5-10) shear strain. We found that fault permeability is highly sensitive to clay content, with permeabilities spanning 2-4 orders of magnitude under otherwise identical conditions. Permeability decreased up to 1 order of magnitude as $\sigma_n$ increased from 2-6 MPa and up to 2 orders of magnitude with increasing shear strain. During shear, halite deformed brittlely, participating in force chains with quartz. Halite and quartz tended to be interchangeable for fault permeability, albeit with slightly lower permeability for halite than quartz, due to porosity reduction via pressure solution and ductile deformation. Our results are applicable to marine-sediment faults with intermixed halite at higher slip rates and/or low effective stress, and have important implications for petroleum reservoir trap assessment in marine-sediment basins.
Introduction

Faults represent significant geologic features for subsurface fluid flow and are important in numerous transport processes within sedimentary basins where they affect groundwater migration [López and Smith, 1995; Evans et al., 1997], petroleum reservoir quality [Smith, 1966], and hydrocarbon production. Fault permeability and poromechanical properties can impact deformation style in subduction megathrust systems [Ikari et al., 2009b]. In particular, faults can act as fluid conduits in low permeability host rock (e.g., granites, basalts, etc.), fluid barriers in permeable matrix (e.g., sandstones), and they may act to redirect flow along their length, as some faults develop significant permeability anisotropy [Evans et al., 1997; Bos and Spiers, 2000; Bos et al, 2000a].

Fault permeability ($k$) can be reduced by up to five orders of magnitude relative to the wall rock, especially in marine sediment environments where clay is integrated into fault zones [Evans et al., 1997; Crawford et al., 2008; Ikari et al., 2009b]. In these environments, faults juxtapose and mix clays from shale along with sands and silts from other horizons [Faulkner et al., 2010]; through abrasional mixing and clay smearing, clay particles distribute through the fault zone, clogging pores between quartz grains. In general fault permeability is thought to be lowest at a critical porosity, depending on mineralogy and development of shear fabric [Revil et al., 2002; Takahashi et al., 2007; Crawford et al., 2008; Ikari et al., 2009b; Niemeijer et al., 2010]. Indeed, small increases in clay concentrations can dramatically reduce fault permeability in sandy faults [Revil et al., 2002; Crawford et al., 2008; Faulkner et al., 2010]. For example, Crawford et al. observed four orders of magnitude permeability reduction as clay content increased from 0 to 50% under otherwise identical conditions. While much work has been done to quantify the role of clay content, load, and strain on fault permeability [e.g., Revil et al., 2002; Crawford et al., 2008], more complex mixtures and the role of ductile minerals such as halite has received less attention, despite the relevance of such mixtures to the salt domes and surrounding country rock.

Impermeable salt domes [Peach and Spiers, 1996] often have associated faults, and these features can act in concert to seal reservoirs [Rowan et al., 1999; Hudec and
In these cases, the fault zones may integrate halite from the salt dome as well as nearby salt layers. Halite undergoes a range of deformation behaviors depending on the chemistry, composition, materials it mixes with, stress, and strain history on the fault, and thus its role for fault zone permeability is complex. Under fast slip conditions halite tends to deform brittlely \cite{Shimamoto, 1986; Bos et al., 2000b; Niemeijer et al., 2010} and should behave as a framework grain. During quiescent, interseismic periods or with slow deformation, halite likely accommodates shear via ductile deformation and/or pressure solution \cite{Rutter, 1983; Shimamoto, 1986; Bos and Spiers, 2000; Bos et al., 2000a; Niemeijer et al., 2008; Niemeijer et al., 2010}.

The purpose of this paper is to present results on the permeability and frictional properties of faults as a function of composition, load, and shear strain. We studied synthetic fault material composed of quartz, halite, and illite shale or smectite clay (montmorillonite). We report on frictional properties and across-fault permeability as a function of strain and effective normal stress. We focused primarily on low effective stress, up to 6 MPa, relevant for shallow burial conditions and sediment basins.

**Methods and Materials**

To measure sediment permeability as a function of normal load and strain, we conducted experiments in the double-direct shear configuration under true triaxial stress conditions \cite[e.g.,][]{Ikari et al., 2009a; Samuelson et al., 2009}. The experimental geometry is shown in figure 4-1. To emulate natural faults in marine-sediment basins, we conducted these experiments on mixtures of quartz sand, illite shale or smectite clay (montmorillonite), and halite. For mixture compositions see table 1 and figure 4-2. These mixtures were designed to test changes across the transition between clast-supported and matrix supported gouge, wherein porosity is lowest and permeability changes are largest \cite{Faulkner et al., 2010}. We used commercial silt-sized quartz foundry sand (F110) and Ca-montmorillonite (sieved to <125 µm). Illite shale and pure halite were ground in a rotary mill and sieved to <125 µm. The illite shale is predominantly comprised of illite, with some quartz and minor kaolinite and plagioclase \cite{Ikari et al., 2009a}. These experiments were carried out under conditions where pressure solution and ductile
deformation of halite is operative at low strain rates and during holds, but with brittle
deformation during shear (e.g., 10 μm/s).

In these experiments, the samples were deformed under confining pressure via
confining fluid, with normal stress and shear stress applied to the sample directly via
hydraulic pistons. These experiments were conducted under double-direct shear, with
normal stress (σn) applied by the horizontal piston, sandwiching the two fault-gouge
layers between three steel blocks. Shear displacement was imposed to the center block by
the vertical piston, inducing shear within the fault gouge layers. Figure 4-3 shows
representative shear stress curves for three experiments. Upstream and downstream pore
pressures (PpA and PpB, respectively) were applied with saturated NaCl brine across the
sample interface, with the hydraulic gradient parallel to σn, and confining pressure was
applied to the entire sample assembly (Figure 4-1). Fast-acting servo-hydraulic
controllers maintained specified horizontal load, vertical displacement, and fluid
pressures. Strain gauge load cells, accurate to +/- 0.1 kPa, measured vertical and
horizontal load. Direct current displacement transducers, accurate to +/- 0.1 μm,
measured vertical and horizontal displacement (shear displacement and layer thickness,
respectively). Pressure transducers, accurate to ± 7 kPa, measured the confining and pore
pressure. Stresses, pressures, and displacements were digitally recorded at 10 kHz with a
24 bit system, and were averaged to 10-100 Hz for storage.

Sample layers were built to specific initial thicknesses from 5 to 7 mm. Porous
steel frits act as the interface between the steel blocks and the sample layers, distributing
the pore pressure and allowing flow parallel to normal stress (Figure 4-1C). The sample
assemblies were sealed with latex jackets, to isolate the sample from the confining oil
(Figure 4-1D). Pore pressures were applied to the sample through the steel blocks, with
upstream pressure (PpA) applied through the center block to both sample layers, and
downstream pressure (PpB) applied equally to the two sideblocks.

Standard experimental protocol was used for each experiment. The sample
assemblies were loaded into the vessel and small initial load was applied (0.6 MPa σn). At
this point layer thickness was measured under load with calipers, and thickness
changes were tracked with a direct-current displacement transducer. The vessel was then
filled with oil, and confining pressure was applied (0.35 MPa). To saturate the sample initially we removed as much air from the sample as possible by pulling vacuum from $P_{PB}$. After ~ 1 minute, pore fluid was introduced through $P_{PA}$ at constant, low flow rate. Flow was maintained until the effluent was free of air for 20 minutes (~ 40 minutes total). After initial saturation, $\sigma_n'$ was increased to 2 MPa (see Table 2 for descriptions of $\sigma_n$, pore pressure and confining pressures). Effective normal stress was calculated following $\sigma_n' = \sigma_n - P_P + 0.5P_C$, where $P_P$ is the average pore pressure, $P_C$ is the confining pressure, and 0.5 is the ratio of $\sigma_n$-parallel contact area exposed to $P_C$. Figure 4-4 shows the standard loading procedure for a representative experiment. For standard loading protocols, $P_{PA}$ applied constant specified pressure to the sample and $P_{PB}$ was held at a constant displacement. Following the permeability test at 2 MPa $\sigma_n'$, load was increased to 4 MPa, and again to 6 MPa, with a $k$ test at each new load. Figure 4-5 shows a representative permeability test. At 6 MPa the sample was sheared 10 mm, after which a $k$ test was conducted. Initial shear was corrected for the elasticity of the latex jacket following standard procedure [Samuelson et al., 2009; Ikari et al., 2009a], accommodating ~ 5 mm of the initial vertical displacement. The sample was again sheared 10 mm and the final $k$ test was conducted. For a representative experiment, figure 4-6 shows the permeability tests at 2, 4, and 6 MPa $\sigma_n'$ with zero strain, and at low and high strain at 6 MPa $\sigma_n'$ (i.e. Perm. $A - E$). Under each of these conditions and prior to each $k$ test, the sample was allowed to compact until compaction rates were less than 0.04 $\mu$m/s (Figure 4-4).

Permeability tests were typically carried out with constant flow rate (Figure 4-5). During these tests, $P_{PB}$ controlled $Q$, and $P_{PA}$ maintained constant pressure, causing a pressure decrease to $P_{PB}$. We measured permeability following Darcy’s Law, $k = Qv h / (A [P_{PA} - P_{PB}])$, where $Q$ is the volumetric flow rate, $v$ is the fluid viscosity, $h$ is the layer thickness, $A$ is the sample contact area (~ 0.005 m$^2$), and $[P_{PA} - P_{PB}]$ is the pressure difference across the sample layers. Steady state flow was not always achieved, with up to ~50% flow rate difference between $P_{PA}$ and $P_{PB}$, likely owing to storativity effects during short-duration tests. Permeability was calculated using both the $P_{PB}$ and $P_{PA}$ flow rates, accounted for in the maximum and minimum reported permeabilities.
Additional error was derived from some pressure fluctuations (e.g., Figures 4-6) and we used peak and trough values to estimate error. Figure 4-6 shows linear flow rates \( q \) and the differential pressures during each of the five permeability tests in p3911, where \( q = Q / A \). To limit dissolution, we generally used very low, constant flow rates \( q < 50 \mu m/s \) and flow volumes, and pore fluid was typically flowed in and out of the sample (e.g., Figure 4-4, sawtooth pattern). These steps limited fresh pore water interaction with the sample. In some of the early experiments, permeability tests were carried out under constant pressure conditions, which generally required larger pressure differences and flow rates, and may have caused minor halite dissolution.

### Results

These marine-sediment fault analog experiments exhibit permeabilities from \( \sim 3 \times 10^{-19} \) to \( > 2 \times 10^{-14} \text{ m}^2 \), within range for this apparatus [Ikari et al., 2009a; Samuelson et al., 2009]. Figure 4-6 highlights the permeability history for p3911, with permeabilities from \( \sim 4 \times 10^{-15} \) to \( \sim 8 \times 10^{-17} \text{ m}^2 \). Figures 4-7 and 4-8 compare the permeabilities during each experiment as a function of load and strain for illite and smectite bearing experiments, respectively (see table 4-1 for error). The highest permeabilities occurred for pure quartz sand, and these values exceeded the permeability limit of the system (\( \sim 2 \times 10^{-14} \text{ m}^2 \)), however previously reported values for quartz permeability are in this range [Zhang and Tullis, 1998]. In all other experiments, the permeability was generally an order of magnitude less than this limit and tended to decrease with clay content (Figure 4-9), as well with increased load and strain (Figure 4-10).

In general, permeability decreased as a function of decreasing quartz content and increasing matrix content (e.g., clay and halite). Figure 4-9 shows permeability contours interpolated between a number of our experiments, prior to and post strain. From 75% to 33% quartz content the permeability is decreased by \( \sim 1 \) to 2 orders of magnitude. In both the smectite and illite suites, the lowest permeabilities occur for mixtures of 33/54/12% by mass quartz/clay/halite, closely followed by the 33/33/33% mixtures. Superimposed on these overarching permeability values are trends of permeability decrease with load and strain.
Permeability decreased across these experiments as a function of normal stress and shear strain (Figures 4-6B, 4-7, and 4-8). With increasing load from 2-6 MPa $\sigma_n'$, all materials experienced 0-1 orders of magnitude permeabilities decrease. Similarly these experiments exhibited 0-2 orders of magnitude permeability decrease with initial strain (from 0 to 2-5 $\gamma$). At higher strains however (5-10 $\gamma$), the samples experienced a range of permeability changes from slight permeability loss to significant permeability increase, up to 1 order of magnitude.

With increasing effective stress, permeability decrease was generally insensitive to quartz content, matrix composition (clay/halite ranging from 5/95% to 81/19%), and clay type (illite vs. smectite; Figure 4-10a-a', respectively). Figure 4-10 shows permeability change due to load and strain as a function of quartz content and matrix composition for all experiments. With initial shear strain permeabilities decreased, and this decrease appears to be slightly enhanced in smectite experiments compared to illite (Figure 4-10B' vs. 4-10B, respectively). Additionally, $k$ decrease tended to be diminished with increasing quartz content, especially in the smectite experiments – for example permeability changed by ~ 1-2 orders of magnitude with < 50% quartz, but only decreased by ~ 0-1 orders of magnitude for higher quartz contents (Figure 4-10). At higher strains, $k$ generally increased and tended to exhibit the opposite trend, with the greatest $k$ increase occurring with the lowest quartz contents. Overall, illite exhibited less permeability change with initial strain (Figure 4-10B), but less permeability increase – and in some cases decrease – with larger strains (Figure 4-10C). We suspect that this relationship owes primarily to the amount of strain experienced by illite versus smectite experiments, since smectite generally attained higher strains over each shear period. Increased shear strain occurred in smectite experiments as a result of enhanced compaction and thus thinner layers.

To study the role of quartz for permeability we compared six experiments with varied quartz content but constant clay to halite ratio (i.e. 50% smectite, 50% halite; Figure 4-11 inset). Figures 4-11B and 4-11C highlight $k$ as a function of quartz content at different loads and strains, respectively. At the highest loads (6 MPa $\sigma_n'$), permeability is lowest at 33% quartz (~ $5 \times 10^{-17}$ m$^2$), and highest at 100% quartz (> $2 \times 10^{-14}$ m$^2$). The
experiment composed entirely of halite and clay showed slightly higher permeability than 33% quartz (~ $3 \times 10^{16}$ m$^2$). Additionally, permeability decreased by 0-1 orders of magnitude with increased load, regardless of quartz content (from 2-6 MPa $\sigma_n$; Figure 4-11B).

Permeability is highest for quartz-dominated mixtures following shear, and in the parametric comparison $k$ was lowest for the mixture with 33% quartz, 33% smectite, and 33% halite by mass (Figure 4-11C). With no quartz content $k$ increased slightly, and with increased quartz content to 100% $k$ increased dramatically. Permeability apparently increased smoothly with increasing quartz content, from 33 to 100% quartz (Figure 4-10C). As identified previously, the lowest permeabilities for most of these mixtures occurs after initial shear strain (2-4 $\gamma$), followed by slight $k$ increase with further strain.

We note that p3976 falls in better line with the other parametric experiments following strain (Figure 4-11C), as opposed to measurements under simple compaction (Figure 4-10B). In this case, shear may have acted as a run-in, distributing particles throughout the matrix in a consistent way, as opposed to any artificial fabric from building the sample [Anthony and Marone, 2005; Chapter 2].

Like permeability, fault friction is strongly dependent on quartz content. Fault layers dominated by quartz (at or above 57% quartz; Figure 4-11A) exhibit similar frictional strengths, at or above 0.6 coefficient of friction ($\mu = \tau / \sigma_n$). With higher matrix content however (≤33% quartz), the fault gouge never exceeds $\mu$ greater than 0.5, and the fault gouge continuously weakens with strain. Weakening behaviors such as these indicate matrix-dominated sliding, perhaps with fabric formation and clay-particle alignment [Rutter et al., 1986; Logan et al., 1992; Haines et al., 2013]. Figure 4-12 shows SEM micrographs of these experiments, with higher magnification in figure 4-13. Fabric formation is particularly evident for the clay rich experiment, p3821, highlighted in figure 4-14. Friction values for higher quartz content indicate frictional sliding between clasts. In particular, frictional sliding with only 57% quartz indicates that halite is participating in clast-type deformation during shear, as brittle deformation of halite may be operative at these load (6 MPa $\sigma_n$) and strain-rate (10 $\mu$m/s) conditions (Shimamoto, 1986; Chapter 2). Additionally, the highest frictional strengths occur with
some matrix content (e.g., 75-57% quartz), indicating that material in the pore space can obstruct efficient frictional deformation [Kaproth et al., 2010]. Peak stress values following holds are likely large owing to frictional healing via halite [Rabinowicz, 1952; Bos et al., 2000b; Niemeijer et al., 2008], and these peaks indicate that the framework clasts are in direct contact with one another, since clay veneer tends to limit fault healing [Dewers and Ortoleva, 1991; Bos and Spiers, 2000].

Halite appears to have multiple roles in affecting the permeability and mechanical properties of the fault zones. It deforms via ductile deformation and pressure-solution [e.g., Bos et al., 2000b] and also via cracking and possibly brittle mechanisms. Halite grains may participate in force chains (Figure 4-13) that carry stress in clast-supported materials during shear [Daniels and Hayman, 2008], indicating brittle deformation. However, halite also shows characteristics of ductile flow, as indicated by grain textures and boundary welding, which is presumably a result of pressure solution [Rutter, 1983; Bos et al., 2000b; Niemeijer et al., 2008]. We conducted permeability tests after the sample had compacted, and thus halite may have acted as a matrix material, between quartz clasts, filling in pore space. In some cases, halite grains mix with clay particles to form localized shear zones. However, micrographs also document halite grains in stress shadows of clast networks and force chains (Figure 4-13). This indicates that halite is not simply a ‘matrix’ material that flows to fill pore space, but rather it can remain in stress shadows and deform brittlely, both of which tend to result in higher pore space than suggested by some previous models [e.g., Revil et al., 2002].

**Discussion**

Throughout our experiments permeability decreases dramatically with small concentrations of clay and halite mixed into quartz fault gouge (Figures 4-9 and 4-11). Overall our permeability values range from a minimum of 5*10^{-19} m^2 to the system limit around 2*10^{-14} m^2. This included a range of strains and loads for quartz/clay/halite end-member mixtures of 100/0/0%, 33/54/12%, and 5/5/90%, respectively. These values fall well-within previously reported values for pure quartz (5*10^{-14} m^2 max. at 25 MPa; Zhang and Tullis, 1998), illite (6*10^{-20} m^2 max. at 40 MPa; Morrow et al., 1984),
montmorillonite-smectite ($7 \times 10^{-22}$ m$^2$ at 40 MPa; Morrow et al., 1984), and halite ($\leq 10^{-21}$ m$^2$ at 5 MPa; Peach and Spiers, 1996), and are in general agreement with previous works on similar granular materials [Crawford et al., 2008; Ikari et al., 2009a]. For example, Ikari et al. reported $2 \times 10^{-16}$ m$^2$ permeability for a 50% quartz, 50% montmorillonite mixture ($\sigma_n' = 12$ MPa), akin to our value of $8 \times 10^{-17}$ m$^2$ ($\sigma_n' = 6$ MPa) for a similar mixture (33% quartz, 54% montmorillonite and 12% halite).

With increased effective stress from 2 to 6 MPa, we observe up to 1 order of magnitude permeability reduction across all of our materials (Figures 4-7, 4-8, and 4-10). Previous studies show similar magnitudes of $k$ reduction, especially at such low normal loads, where anelastic compaction is dominant [Morrow et al., 1984; Faulkner and Rutter, 2003; Crawford et al., 2008; Ikari et al., 2009a]. Permeability reduction was generally indiscriminate of the mixture composition, as all materials underwent compaction with increased load. Under simple compaction, it is likely that intact natural materials will have lower permeabilities, as naturally consolidated materials tend to be more compact than remolded materials. Additionally, morphology differences between natural clay and our crushed materials may result in systematic permeability differences [e.g., Ikari et al., 2009a].

With shear, fault permeability decreased by up to 2 orders of magnitude, and tended to be enhanced in clay-rich materials (Figures 4-7, 4-8, 4-10, and 4-11). These changes are in general agreement with previous laboratory works on synthetic fault zones [Morrow et al., 1984; Zhang and Tullis, 1998; Crawford et al., 2008; Ikari et al., 2009a]. Shear driven compaction, fabric formation, and abrasional mixing all tend to assist $k$ reduction with fault development [Faulkner et al., 2010]. Abrasional mixing is especially important in nature because it allows comingling of different wall rock materials, mixing clay in with clasts. Similar processes likely occur during our experiments, mixing the gouge and erasing any fabric artificially made during the sample build (e.g., Chapter 2). This may help to explain why p3976 falls into trend with other experiments as a function of quartz content in figure 4-11C following shear.

Throughout these experiments, halite particles show characteristics of both brittle and ductile deformation. Some halite grains appear to participate in force chains, which
carry stress in a clast-supported gouge [Daniels and Hayman, 2008], and show some
evidence of grain cracking. However, halite grains also show evidence of ductile
def ormation, such as bending around quartz and halite clasts, and pressure solution. In
 particular, figures 4-13 and 4-14 show halite grains welded to quartz particles and clay
 fabric, evidence for pressure solution [e.g., Bos et al., 2000b]. We suspect that some
halite grains deform brittlely during shear (10 µm/s), resulting in clast supported
frictional behavior with low quartz concentrations (e.g., 33% quartz; Figure 4-11).
However, ductile deformation and pressure solution likely predominate during holds, and
these processes may also occur for halite grains sitting in stress shadows during shear.

Although halite acts as matrix during permeability tests, it appears that relicts of
shear-driven brittle deformation are maintained during holds and $k$ tests. SEM
micrographs shows some halite grains involved in force chains, yet these thin-sections
were made following a long hold (~ 30 minutes) leading to the final permeability test
(Figure 4-13). Additionally, our parametric analysis shows that $k$ was lowest for the
mixture with 33% quartz, 33% halite, and 33% smectite by mass, or 28% quartz and 72%
matrix by volume, but previous models suggest that $k$ should be lowest with 25-40
volume % matrix [Revil et al., 2002; Crawford et al., 2008]. These observations indicate
that halite does not simply participate as matrix, flowing between quartz clasts and
efficiently infilling pore space. Indeed, figure 4-13A-B shows groupings of halite grains
with intermixed clay, leaving some additional pore space between quartz grains. Instead,
minimum porosity, and perhaps minimum $k$, likely occur with some higher clay content,
making up for the non-ideal distribution of halite grains.

Experiment p3909 was matrix dominated and behaved as such during shear and
permeability tests. Figures 4-12 and 4-13 show that p3909 develops fabric within the
matrix, with clay and halite aligning in the P orientation. This material also weakened
significantly during shear, highlighting fabric generation within the matrix-supported
gouge. For permeability in this case, fluid flow through matrix and is partially retarded by
the impermeable quartz grains. We suspect however that $k$ might be lower with a slightly
higher clast concentration, at the threshold between a matrix and clast supported gouge.
Additionally, we expect that for a given concentration of clay, $k$ will be preferentially
lower with halite over quartz since halite can infill some neighboring pore space via pressure solution [Dewers and Ortoleva, 1991; Niemeijer et al., 2008] and ductile deformation during holds, bonding with quartz grains and clay fabric (Figure 4-9). Additionally, pressure solution is likely enhanced during holds by the presence of clay particles, which tends to increase diffusion rates [Dewers and Ortoleva, 1991; Renard et al., 2001].

The permeability implications for brittle halite deformation during shear and ductile deformation and pressure solution during holds is likely relevant for coseismic faults or faults with very low effective stress [Shimamoto, 1986; Davison, 2009]. At slower slip rates (e.g., aseismic faults), higher effective stresses, or higher temperatures, halite may accommodate fault slip via ductile deformation and pressure solution. With matrix-like behavior at all times, pore space should be efficiently infilled from 60-75 volume % quartz [Revil et al., 2002]. However, we suspect that permeability would be lowest for near-pure mixtures of halite and clay, as Bos and Spiers [2000] showed that these mixtures develop mylonite-like fabric, with halite particles stretching parallel to the fault. Since halite grains are near impermeable [Peach and Spiers, 1996], mylonite fabric might dramatically reduce fault perpendicular flow.

In our matrix supported materials, non-mylonitic fabric is generated within the matrix material, which may affect gouge permeability [Faulkner et al., 2010]. Fabric generation appears to be particularly enhanced with greater clay content (Figure 4-14 vs. Figure 4-13). Fabric development occurs with rotation and preferential alignment of clay particles into the P-orientation [Haines et al., 2013], near perpendicular to our flow direction, perhaps decreasing permeability by up to 1 order of magnitude [Faulkner et al., 2010]. With greater shear however, macro-fabric features (e.g., Riedel, P, and Y shears) are developed [Logan, 1992; Haines et al., 2009; Haines et al., 2013], and fluid flow may be enhanced along these features [Arch and Maltman, 1990; Zhang and Tullis, 1998; Zhang et al., 1999]. This may be especially true along high angle P-shears, which readily develop in high clay content experiments (e.g., p3821; Figure 4-14). In this way, $k$ increases in clay rich fault gouge from low to high strain (Figure 4-10), similar to observations by Crawford et al., [2008]. Previous studies at higher loads show continued
\( k \) decrease with strain, likely owing to the effects of comminution and compaction [Ikari et al., 2009a; Faulkner et al., 2010], which was limited in our low-stress, quartz-rich system, noting the intact grains in figure 4-13.

**Conclusions**

These experiments were designed to investigate the permeability evolution of marine sediment faults, as well as the potential role of intermixed halite. Through these experiments we found that fault permeability is highly dependent on clay content, effecting baseline \( k \) by up to two orders of magnitude under otherwise identical conditions. Additionally, we found up to 1 order of magnitude permeability decrease with increased load (\(< 6 \) MPa \( \sigma_n \)) and 2 orders of magnitude decrease with small strain (\( \gamma < 5 \)). These results are in general agreement with previous works showing the effect of compaction with load and shear driven abrasional mixing and clay smearing for fault permeability [Morrow et al., 1984; Zhang and Tullis, 1998; Faulkner and Rutter, 2003; Crawford et al., 2008; Ikari et al., 2009a].

These experiments are analogous to coseismic faults, or faults that have periods of relatively fast slip, with periods of quiescence, as halite gains show evidence for brittle deformation during shear [Shimamoto, 1986; Davison, 2009]. Although halite shows matrix-like behaviors during holds and leading up to permeability tests, including pressure solution and ductile deformation, we find that the general geometry of particles following shear tends to remain intact. As a result, halite and quartz are generally interchangeable for fault permeability, albeit with slightly enhanced \( k \) reduction for halite, owing to enhanced compaction during holds. We find it likely that halite would play a different role with aseismic fault slip, with the potential to develop mylonitic fabric with clay particles, severely limiting fault permeability [Bos and Spiers, 2000; Bos et al., 2000a]. Our results suggest that to assess marine sediment faults for reservoir trap viability and fluid flow, one must consider gouge composition, load, strain, and strain rates.
Figure 4-1. Experimental setup. Experiments were conducted in a biaxial stress apparatus with a true triaxial pressure vessel (A). The double direct shear assembly was placed in the pressure vessel, and confining pressure, $P_C$, pore pressure inflow, $P_{PA}$, and pore pressure outflow, $P_{PB}$ were applied through steel lines to the sample blocks (B). Normal stress was applied horizontally across both sample layers, and shear stress was applied vertically to the center block, inducing shear within the sample layers (C). Pore fluid access to sample layers was via channels and porous metal frits in the side blocks (D). The sample and pore pressures are isolated from the confining pressure with latex rubber jackets.
Figure 4-2. Ternary diagram showing synthetic fault gouge compositions. Experiments were carried out on mixtures of quartz, halite, and clay (illite shale or montmorillonite). A total of 18 experiments were conducted. Symbol color and experiment number denote mixture composition here and in subsequent figures.
Figure 4-3. Friction curves from three representative experiments. See Figure 2 for composition details. Each curve shows elastic loading and the onset of plastic strain and shear deformation upon initial loading. After peak stress, materials with high clay content weaken dramatically, and, for the case of no quartz, a coefficient of friction, \( \mu \), near 0.3 is reached. Frictional strength increases with increasing quartz content; maximum residual friction values \( \mu \) for steady sliding are 0.6-0.7, which is typical for a clast-supported mixture. In each experiment shear was stopped (hold periods) during permeability tests and then resumed, resulting in a drop in shear stress followed by a large peak that represents sample lithification and frictional healing.
Figure 4-4. Complete history for a representative experiment. Normal stress is applied initially to 0.6 MPa and then confining pressure is applied followed by pore pressure. Note history of effective normal stress $\sigma'_n$ shown below Pore Pressure A in the upstream reservoir (Figure 1). After 1 hour of initial fluid saturation, sample were loaded to 2 MPa $\sigma'_n$ through a combination of increased pore pressure, $\sigma_n$, and confining pressure, $P_C$. A permeability test (i.e. Perm. A) was conducted at constant flow rate after the sample reached steady-state compaction. Following the perm test, the sample was loaded to 4, and 6 MPa, and then the sample was sheared. Each of these steps preceded a permeability test.
Figure 4-5. Representative permeability test with constant flow rate. During each test a constant flow rate was applied at $P_B$, driving a differential pressure between $P_A$ and $P_B$. We measure permeability once flow reaches steady-state, with uncertainty given by noise and small fluctuations in differential pressure.
Figure 4-6. Details of each permeability test from one experiment, p3911 (See Figure 4). Panel A shows the linear flow rate, $q = Q/A$, and resultant differential pore pressure for each $k$ test. With increasing load and shear strain, we decreased $q$ to limit $\Delta P$, as $k$ tended to decrease with each step. Low flow rates were used to limit halite dissolution and momentum transfer effects between fluid and grains. Panel B shows the change in permeability as a function of load (left side) and shear strain (right side), highlighting the tendency for $k$ decrease with load and strain.
In general, $k$ increased with high quartz content, and baseline permeabilities spanned $\sim 2$ orders of magnitude, $OM$. For 100% quartz, p3979, we measured very high $k$, exceeding the system measurement limit ($k \sim 2\times10^{-14}$ m$^2$). Across all experiments $k$ decreased 0-1 $OM$ with load. Similarly, $k$ decreased with low-level strain by 0-2 $OM$, and was enhanced with high clay contents. At higher strains however, $k$ tended to level out or increase slightly, as was typically the case for high matrix content experiments at very high strains ($\gamma > 8$).
Permeability increased with increasing quartz content, with baseline permeabilities spanning ~ 2.5 $10^{-14}$ m$^2$. $k$ decreased with increasing load by 0-1 $10^{-14}$ m$^2$. $k$ decreased by 0-2 $10^{-14}$ m$^2$ as shear strain increased from 0 to ~3; the effect was greater for the samples with the highest clay content. $k$ did not change dramatically at higher shear strains, and tended to increase in matrix-rich experiments at higher strains ($\gamma > 6$).

**Figure 4-8.** Permeability as a function of load and strain for illite shale experiments. Permeability increased with increasing quartz content, with baseline permeabilities spanning ~ 2.5 $10^{-14}$ m$^2$. $k$ decreased with increasing load by 0-1 $10^{-14}$ m$^2$. $k$ decreased by 0-2 $10^{-14}$ m$^2$ as shear strain increased from 0 to ~3; the effect was greater for the samples with the highest clay content. $k$ did not change dramatically at higher shear strains, and tended to increase in matrix-rich experiments at higher strains ($\gamma > 6$).
Figure 4-9. Permeability as a function of mixture composition interpolated between most experiments (illite on left panels, smectite on right panels experiments), at 6 MPa with no strain (Panels A, A’) and at 6 MPa with low strain (Panels B, B’). Permeability tended to be decreased with increased clay content before and after strain. $k$ tended to be lower for halite relative to quartz at constant clay content. Note that permeability does not show systematic differences between the illite and smectite clay experiments.
Figure 4-10. Trends of permeability loss as a function of load (A and A’), low strain (B and B’), and high strain (C and C’) for illite and smectite samples, respectively. Colors denote matrix composition (See Figure 1). With increased load, $k$ tended to decrease by 0-1 orders of magnitude, with little dependence on clay type (A vs. A’), quartz content, or matrix composition. With low strain, permeability tended to decrease by 0-2 $OM$, with the strongest change occurring for clay-rich experiments. With increased strain however, $k$ changed little, with the strongest increases occurring for experiments with the highest final strains (Table 4-1) and clay contents.
Figure 4-11. Fault properties as a function of quartz content, with constant clay to halite ratio. These experiments show an evolution from high to low frictional strength with decreased quartz content (A). In particular, matrix dominated materials weaken dramatically with shear strain, likely resulting from fabric development [Haines et al., 2013], and quartz dominated materials maintain friction values above 0.6. Quartz dominated materials with some matrix exhibit more healing, likely resulting from the halite content [Niemeijer et al., 2008; Chapter 2], and generally higher stresses, resulting from pore-space infilling [Kaproth et al., 2010]. Panels B and C show the permeability evolution with quartz content during these experiments as a function of load and strain, respectively. In general, increased quartz content results in increased $k$, with an apparently smooth trend from 100% to 33% quartz. Permeability appears to increase slightly from 33% to 0% quartz. Experiment p3976 is slightly off trend with the other experiments, but after small strain comes into line with other experiments, indicating the effects of run-in shear to develop a consistent matrix [Chapter 2].
A  p3908 – 75% Qtz, 12.5% Smectite, 12.5% Halite

B  p3976 – 66% Qtz, 17% Smectite, 17% Halite

C  p3910 – 57% Qtz, 21% Smectite, 21% Halite

D  p3909 – 33% Qtz, 33% Smectite, 33% Halite

E  p3978 – 0% Qtz, 50% Smectite, 50% Halite
Figure 4-12. SEM micrographs from five experiments with varied quartz content and constant halite to clay ratio (e.g., Figure 4-11). Each of these thin sections was made following the final permeability test, and are thus at high strain (Table 4-1). Halite particles are bright-white, quartz particles are large and gray, and the gray matrix is smectite clay (individual particles are not visible, at < 1 µm diameter). At high quartz content the faults are framework grain supported (A-C), but are matrix supported at lower quartz content (D-E). Fabric generation appears to only occur in the matrix-supported gouges. Although salt acts as matrix during holds, during shear it appears to be load-bearing, acting as part of the grain-framework skeleton. In particular, p3910 shows halite grains apparently within force chains with quartz grains. Since these thin sections were made following the final hold, salt grains show strong evidence plastic and pressure solution behavior, which are highlighted in figure 4-13.
Figure 4-13. Higher magnification SEM micrographs from figure 4-12. High quartz experiments do not develop a strong matrix fabric (A-C, C’), but fabric is well developed with high matrix content (D-E). Most pore space, especially along quartz grain boundaries, is open as a result of smectite dehydration and densification following each experiment. However some of this pore
space was likely open while the sample was hydrated and intact. In many of these experiments (e.g., p3976) halite grains appear to be in brittle contact with other halite or quartz grains, indicated by $B$. This allows for accumulation of clay particles between halite grains, potentially leaving pore-space between quartz grains, indicated by $\phi$. Halite welding (via pressure solution) is evident throughout these micrographs, noted as $W$, and in some cases the halite appears to flow with the clay fabric, noted as $P$. Panels C and F highlight sister experiments with similar mixture ratios, but contain smectite (montmorillonite) clay and illite shale, respectively. These mixtures are similar, but the matrix appears more granular with illite than smectite, owing to the contrast in particle size and some quartz content of the illite shale.

Figure 4-14. SEM micrograph from smectite dominated experiment, p3821 (54% smectite, 33% quartz, 12% halite). Like the other experiments, halite welds with other grains ($W$) and is in brittle contact with some quartz grains ($B$). P-shears developed throughout this material, typical of clay-rich materials. Additionally, halite grains flow with and weld to the clay fabric, and apparently do not locally disrupt the clay fabric, unlike quartz grains.
Table 4-1. Permeability measurements for all experiments.

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Appendix A: Details for all experiments

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<td>10-30</td>
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<td>Chap</td>
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<td>Material(s)</td>
<td>Vsl.</td>
<td>AC</td>
<td>Sat.</td>
<td>$\sigma_n$ (MPa)</td>
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Appendix B: Code for acoustic data reduction and porosity modeling

i: A_listgrab (MatLab m files)

%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%
%%% Code designed to take all accoustics files and make a list of their
%%% names, sorted by the time they were made. This code will also take in
%%% experimental data and match it to accoustics data of the same time.
%%% Uses a voltage ramp to synch between acoustic files and biax recorder.
%%% Voltage is appended to name of every AC file, this pulls off that
%%% voltage
%%% B. Kaproth 8.6.12
%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%

clear all
close all

name='p3917';
delay = 0; %Time delay between acoustics and computer, may need to adjust to
get windowing of voltage to align

files = dir(fullfile('/Volumes/DATA/',name,'AC')); %folder with acoustics data should be
e.g., p3917AC
names=struct2cell(files'); %puts each files info into a matrix with rows of
name, time, bytes, etc
names=names(3:end,1:2);
names=sortrows(names,2); %sort based on time of file creation
time=datevec(names(:,2)); %switches to a date vector system
c=1;
S='log';
difference=etime(time(1,:),time(length(time),:))-delay;

while(c<=length(names)) %assigns a number if the file is a log file (1) or a
different file (0)
    k=strfind(names(c,1),S);
    k=cell2mat(k);
    if(isnan(k)==0)
        comp(c)=1;
    else
        comp(c)=0;
    end
    names{c,2}=etime(time(c,:),time(length(time),:))-difference; %reassign
    experimental time
    c=c+1;
end

comp=comp';
c1=1;
c2=1;
while(c1<=length(names)) % reads through all non-log files and puts them into a matrix
    if(comp(c1)==0) % slick sorter to skip lines of log files
        record(c2,1:2)= names(c1,:);
        tempname= names(c1,1);
        cutoff= strfind(tempname,'_'); % this should find the underscore before the voltage starts
        vol= tempname(cutoff+1:end-4); % this remakes the name string after '_' and without '.txt'
        record(c2,4)= str2num(vol);
        cutoff2= strfind(tempname,'rec'); % find the first number after 'rec' and '_'
        recnum= tempname(cutoff2+3:cutoff-1); % pull the number between 'rec' and '_', should be the record number
        record(c2,3)= str2num(recnum);
    c2=c2+1;
    end
end
c1=c1+1;
end

c4=1; % this is the counter that will cycle its way through 'record'
for c3 = 0:2:record{length(record),2} % stop at every '1 second' interval, i.e. column 2
    leftcnt= c4;
    while record(c4+1,2) == c3 % this pulls all the rows at a given 'second interval' i.e. 0 secs is row 1-31
        c4=c4+1;
    end
    rightcnt= c4; % i.e. leftcnt is row 1 and rightcnt is row 31
    record(leftcnt:rightcnt,:)= sortrows(record(leftcnt:rightcnt,:),3); % sort based on column 3 (rec num), rows 1-31
    c4=c4+1;
end
cd(strcat('/Volumes/DATA/',name));
save record record
**ii: B_listgrab (MatLab m files)**

%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%
%%% Designed to reduce data from experiments before matching to your
%%% acoustic data
%%% rawdata.txt is dumped out of XLOOK, using type_p (no header)
%%% See columns below
%%% written by B. Kaproth
%%% 2.15.2010
%%% 8.6.2012 last update
%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%

```matlab
clear all
close all

name='p3917'; % your filename

cd(strcat('/Volumes/DATA/',name))
load record
load rawdata.txt % this is the name of your dumped data, in ASCII
data(:,1:9)= rawdata(:,1:9);
synch(:,1)= data(:,5); % pulls the time out of the data file
synch(:,2)= round(data(:,9)*10000); % mV*10; Pulls the voltage from volt ramp

% synch
offset= 26.54; % dig this time delay out by finding when acoustics started with respect to the experiment start

for c1 = 201:length(record(:,1));
    time= record(c1,2)+offset; % This is the acoustic file's time
    volt= record(c1,4)+0; % This is the acoustic file's voltage
    while time - 2 > synch(c2,1) % Data file, left time window (time -2s)
        c2=c2+1;
    end
    c3=c2;
    while time + 2 > synch(c3,1) % Data file, right time window (time +2s)
        c3=c3+1;
    end
    if volt > 5000 % when at high voltage, start at left time, and work your way up until voltage satisfied
        c4=c2;
        while volt > synch(c4,2)
            c4=c4+1;
        end
        c5=c4;
        c4=c4-1;
    else % when at low voltage, start at right time, and work your way down until voltage satisfied
        c4=c3;
        while volt < synch(c4,2)
            c4=c4-1;
        end
        c5=c4+1;
    end
```

125
\[ \text{diffbig} = \text{synch}(c5,2) - \text{volt}; \] 
% the AC voltage will split the difference between two recorded points, these next few lines find the closest one

\[ \text{diffsmall} = \text{volt} - \text{synch}(c4,2); \]

\[ \text{if } \text{diffbig} > \text{diffsmall} \]
\[ c4 = c4; \]
\[ \text{else} \]
\[ c4 = c5; \]
\[ \text{end} \]

\[ \text{match}(c1,1) = \text{volt}; \text{match}(c1,2) = \text{synch}(c4,2); \] 
% for a given acoustic file, this shows the match between the AC volt and the data volt

\[ \text{synch}(c4,3) = \text{time}; \text{synch}(c4,4) = \text{volt}; \text{synch}(c4,5) = c1; \] 
% col 3 becomes the AC time, col4 becomes the AC volt, col5 becomes the AC file number

\[ \text{dat}(c1,1) = \text{data}(c4,1); \] 
% vdisp
\[ \text{dat}(c1,2) = \text{data}(c4,2); \] 
% Sstress
\[ \text{dat}(c1,3) = \text{data}(c4,3); \] 
% thicness
\[ \text{dat}(c1,4) = \text{data}(c4,4); \] 
% Nstress
\[ \text{dat}(c1,5) = \text{data}(c4,5); \] 
% time
\[ \text{dat}(c1,6) = \text{data}(c4,6); \] 
% mu
\[ \text{dat}(c1,7) = \text{data}(c4,7); \] 
% strain
\[ \text{dat}(c1,8) = \text{data}(c4,8); \] 
% edisp
\[ \text{dat}(c1,9) = \text{synch}(c4,2); \] 
% data voltage
\[ \text{dat}(c1,10) = \text{volt}; \] 
% AC voltage

\[ \text{end} \]
\[ \text{clear data} \] 
% remove the data file
\[ \text{recdub} = \text{dat}; \]

\[ \text{save recdub recdub} \] 
% recdub is the name of our experiment data matrix (e.g. friction, disp...)
iii: C_Imagemap_k_v4 (MatLab m files)

%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%
%%%File designed to take names file (list of all waveform file names in a
%%%vert matrix), go through these names, and attach the data contained
%%%within each file. Will make a matrix of all waveforms through time.
%%%Created by M Knuth 3/2/09
%%%edited by Kaproth 8/6/12
%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%
clear all
close all
clc
name='p3917';

name=record(:,1);
filenumber = size(names);
filenumber = filenumber(1);
counter=1;

while (counter<=filenumber(1))
a=names{counter};

amplitude=load(a);

if amplitude(1,1) == 1
amplitude = amplitude(3:length(amplitude),1)
elseif amplitude(1,1) == 4096
amplitude = amplitude(3:length(amplitude),1)
end

if length(amplitude)>=length(ampmap(1,:))
ampmap(counter,:)=amplitude(1:lngth);
else
ampmap(counter,1:length(amplitude))=amplitude(:);
end

end
cd(strcat('./../',name))
iv: D_Image_Picker_Oneshot (MatLab m files)

%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%
%File to explore the ampmap file.
%Designed to allow the user to pick multiple
%rows of data from ampmap, plots one at a time

% Use right click to kill the operation
%created by B. Kaproth 4/2009
%edited 8/12
%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%

clear all;
close all;

name='p3917';
prerows=214;%first 214 rows of each AC file is before the wave is sent.

cd(strcat('/Volumes/DATA/',name));
load ampmap;
%load recdub;
firstrow= 1;
lastrow=length(ampmap(:,1));
ampmap=ampmap(firstrow:lastrow,:);
hold off
figure(1)
imagesc(ampmap);
colormap(jet);

%these 5 lines will show you your p-picks, once made, on the ampmap
p=plot(recdub(:,11),1:length(recdub(:,1)),'k')
set(p,'linewidth',2)
axis([214 1500 0 length(recdub(:,1))])
hold off

set(gca,'YDir','normal')
caxis([-0.015 .015])
axis tight
y=0;
button=1;
count=4;

figure(2) %plot time
plot(recdub(firstrow:lastrow,5))
figure(3) %plot shear stress
plot(recdub(firstrow:lastrow,2))

while(button<1.5); %right click kills; every left click will access that row of
ampmap, showing the waveform, zero time, calculated p-arrival, estimated s-
arrival
figure(1)
[x,y,button]=ginput(1);
y=round(y) %need to round to an integer, because rows are integers
figure(count) %assigns every waveform to its own figure
plot(ampmap(y,:)) %picks the waveform of a given row out of ampmap

title(strcat('Row #', num2str(y))); ylabel('amplitude'); xlabel('time, .04
microsec');
    hold on %allows us to plot different points of interest (p-wave, s-wave, zero) on same plot
    plot(prerows,0,'m+') %this is number of rows before the signal is sent, 263
    for settings feb 2010
        %plot p and s arrivals on waveforms, once you have picked them
        plot(recdub(y,9),ampmap(y,round(recdub(y,9))),'rs') %plot the p-arrival
        plot(recdub(y,10),ampmap(y,round(recdub(y,10))),'rs') %plot the p-arrival
        plot(recdub(y,13),ampmap(y,round(recdub(y,13))),'g*')
        plot(recdub(y,8),ampmap(y,round(recdub(y,8))),'g*')
        hold off
        axis([200 1000 -.1 .1]); %wide
        axis([1 1000 -.005 .005]) %zoomed p-wave
        count=(count+1);
    end
v: E_Xcorr_s_fast2 (MatLab m files)

%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%
% cross correlation picker
% Run separately for S and P waves
% %
% Pick a master wave -- 1 full cycle (2*pi) -- with a very good master wave
% Compare to the other waves
% Since this is a moving window, you will have to reset the window position
% multiple times to pick all waves in a given experiment
%
% Created by B. Kaproth 8_2010
% Edited 8/12
%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%

%%

name='p3917';
cd(strcat('/Volumes/DATA/',name));

window=166; %from master wave, this is a half cycle (e.g., window is 166 rows long)
load ampmap %shut this off after you load it first
handpick=630; %the best p-wave handpick for the master wave
xi=0:.1:window; %for interpolation, .1 interpolates 10x
x=0:window;

pmaster=ampmap(1301,handpick:handpick+window); %draw out the master wavelet
pmaster=interp1(x,pmaster,xi,'spline'); %now, actually interp the master wavelet
maxcorr=sum(pmaster.*pmaster);

clear ampmap
%%

clear corr
%name='p3455';
cd(strcat('/Volumes/DATA/',name));
%load ampmap
load recdub;
TooMany = 0;

up = 2; % 1 moves up in row number 100--->101; 2 moves down 100 --->99
TooManyLimit = 300; % will shut the program down if there are too many bad picks (using correlation coefficient)

oldpick=690; % This sets the left hand side of the moving window for all picks. Readjust this to get all picks

sample=512;%set the record you wish to start from

mx = 2000; % run program to this record (limit it so that you can shift the window via 'oldpick'

mn = 500; %Or, run program down to this record

if up == 1;
   while sample<=mx; %sweep through all wave forms
      xi=0:.1:window+40; %40 is window size, interpolate on this
      x=0:window+40; %40 is window size
      pwave=ampmap(sample,oldpick-20:oldpick+window+20); % sample wavelet 20 counts before and after, 20 is half of window size
      pwave=interp1(x,pwave,xi,'spline'); %interpolate
      cnt=1;

      maxcorr=sum(pmaster.*pmaster);
      % Let's see if it's a good pick
      if maxcorr>TooManyLimit
         % It is, so we keep it and move the window
         corr=corr+1;
         oldpick=sample;
         oldpick=oldpick-40;
         sample=sample+40;
      else
         % It's not a good pick, so we throw it away and move the window
         corr=corr-1;
         oldpick=sample;
         oldpick=oldpick-40;
         sample=sample+40;
      end
   end
end
maxcnt = length(pwave) - length(pmaster);
while cnt < maxcnt % sweep through the sample with master wave; shift
master wave along wavelet of interest
    corr(1,cnt)=sum((pmaster).*(pwave(cnt:cnt+window*10)))); % multiply
waves at the current phase shift, cnt.
    cnt=cnt+1;
end
[trash,row]=max(corr); % find the max correlation of the two waves, this
is our pick
wavecorr=sum(pwave(row:row+window*10).*pwave(row:row+window*10));
corrperc(sample)=corr(row)/(wavecorr^(1/2)*maxcorr^(1/2)); % calculates
the correlation coefficient
row=(row-201)/10; % before this, pick is -- 20 rows before *10 --
correct this out
spicks(sample)=oldpick+row;

if corrperc(sample)<.30 && sample< length(recdub(:,1)) % loop to kill
program if too many bad picks
    TooMany = TooMany +1;
    if TooMany>TooManyLimit
        sample = mn
    end
end
sample=sample+1;
end

elseif up == 2;
    while sample>mn;
        xi=0:.1:window+40; % need to be 2x "oldpick-blah:oldpick+blah"
        x=0:window+40;
        pwave=ampmap(sample,oldpick-20:oldpick+window+20); % get sample wavelet
        pwave=interp1(x,pwave,xi,'spline'); % interpolate
        cnt=1;
        maxcnt = length(pwave) - length(pmaster);
        while cnt< maxcnt % sweep through the sample with master wave
            corr(1,cnt)=sum((pmaster).*(pwave(cnt:cnt+window*10)))); % multiply
            waves at the current phase shift, cnt. I cube these, to make high values
            prefferential
            cnt=cnt+1;
        end
        [trash,row]=max(corr); % find the max correlation of the two waves
        wavecorr=sum(pwave(row:row+window*10).*pwave(row:row+window*10));
corrperc(sample)=corr(row)/(wavecorr^(1/2)*maxcorr^(1/2));
        row=(row-201)/10; % before this, pick is -- 20 rows before *10 --
correct this out
        spicks(sample)=oldpick+row;
        if corrperc(sample)<.80 && sample< length(recdub(:,1)) % spicks(sample)=spicks(sample+1);
            TooMany = TooMany +1;
            if TooMany>TooManyLimit
                sample = mn
            end
        end
        sample=sample+1;
    end

end

%%
figure(1)
subplot(3,1,3)
plot(corrperc, '.')
axis tight
ylim([0 100])
xlim([mn mx])

subplot(3,1,2)
plot(spicks, '.')
axis tight
ylim([oldpick-50 oldpick+50])
xlim([mn mx])

%subplot(3,1,1)
%imagesc(ampmap(:,1:1400)')
%caxis([-0.05 .05])

%recdub(:,13)= spicks; recdub(:,14)= corrperc; save recdub recdub %this
%is for coda
%recdub(:,11)= spicks; recdub(:,12)= corrperc; save recdub recdub %this
%is for absolute p-pick
%recdub(:,9)= spicks; recdub(:,10)= corrperc; save recdub recdub %this
%is for absolute p-pick
vi: GeomThinModel (MatLab m files)

%%%%%%%%%%%%%%%%%%%%%%%%%%%%%
% Model to estimate porosity and mass loss throughout friction experiments
% This was used extensively in chapter 2, Kaproth-Gerecht Thesis
% For halite, we found that a shapefactor of 1 (rectangle) best predicts mass loss
% A shape factor of 0.5 is a triangle
% built by B. Kaproth 2.6.2012
%%%%%%%%%%%%%%%%%%%%%%%%%%%%%
clear all
close all

name = 'p3635';
for i = 1:2
    imass = 30.519; % this is the mass at the end of the experiment, we call this initial because it is our best measurement of mass
    cd(strcat('/Volumes/DATA/', name));
    load recdub
    thkerror = 0; % estimate thickness error
    fine = length(recdub(:,5)); % last row
    c = fine;
    rho(c) = imass/(5*5*(recdub(c,3)+thkerror)*2/10000); % set initial density
    xdim = 5; % height of sideblock, ydim falls out, in cm
    ydim = 5;
    masslost(c+1) = 0;
    vollost(c+1) = 0;
    shapefactor = 1;
    ...
    for c = fine : -1 : 2
        thickt = (recdub(c,3)+thkerror) * 2 / 10000; % layer thickness in microns at time t, in cm
        thicknew = (recdub(c-1,3)+thkerror) * 2 / 10000; % layer thickness in the future timestep, in cm
        rhot = rho(c); % rho at time t
        dx = (recdub(c,1) - recdub(c-1,1)) / 10000; % displacement in cm
        rect = (dx * thickt * rhot) * shapefactor;
        rhot = (rhot * thickt * xdim) + rect) / (thicknew * xdim); % Modeled parameter - predict density at the next timestep
        masslost(c) = masslost(c+1) + rect * ydim; % expected mass of the wedge through time
        vollost(c) = vollost(c+1) + dx * thickt * ydim / 2; % expected volume of the wedge through time
        samplemass(c) = rhot * thickt * xdim * ydim; % expected mass of the sample through time
    end
end
samplemass(2)
masslost(2)
vollost(2)
plot(rho)
phi = 1 - rho/2.16;
figure(2)
plot(phi)
recdub(:,20) = rho; recdub(:,21) = phi*100; save recdub recdub;
Appendix C: PZT installation with Faraday cages

Install 0.5” pink diamonite wear plates. Use a thin layer of Epotec epoxy, 730 unfilled. Put the silver coated surface to the top. Check to make sure no electric connection between the surface of the wear plate and the blocks.

Install the PZT on to the top of the silver wear plate. Use silver epoxy (not electrically conductive until it sets). If it is a polarized transducer, dual mode or shear wave, the polarized direction has a chip off of the side. Make sure the polar direction runs parallel with the block teeth.

Install a brass wall around the outside of the transducer. This is made simply with a thin brass sheet ~1cm tall. Using silver epoxy, attach this to the silver wear plate. At this point, we have three electrically distinct parts of these blocks – check to make sure no apparent connections. 1) The steel block. 2) The top of the silver wear plate, with the bottom of the PZT, and with the brass wall. 3) the top of the pzt.

Build a top circle that will fit the top of the brass wall. Punch a hole in this that will fit the signal wire and casing of a coax cable (use a penny nail).
At this point, strip back the braid from the signal wire and its casing. Pull the signal wire out from the braid \(~1-2”\). Stick the signal wire through the whole in the brass top. Spread the braid out over the brass lid and solder it down. Be limited in iron use it is easy to burn through the signal wire.

Completed coaxial cable with endcap.
Attach the signal wire to the top of the pzt with silver epoxy. Cure for one day.

Pull the coax casing down to the PZT. Do this by gripping the signal cable at the end of the coax cable and pulling down on the casing. Do not pull too far, otherwise you will rip the signal cable off of the top of the pzt.

Attach the cap briefly with a few points of solder. Seal up the cap with silver paint. Remember, the top of the pzt should not be in electrical connection with any of the brass case/braid.

Completed setup for double direct shear.
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EDUCATION
Ph.D. in geosciences, Penn State University, August 2013
Advisor: Dr. Chris Marone
Coursework includes: Fault mechanics, Mathematical modeling in the geosciences, Quantitative methods in hydrology, Advanced seismology, Dynamics of the Earth, etc.

BS geology, SUNY Brockport, magna cum laude, 2007

EXPERIENCE
• Research Assistant, Rock & Sed mechanics with Chris Marone, Penn State University, 2007-present
• BP Intern, Geomechanics & Gulf of Mexico well planning, Summer 2011
• Teaching Assistant, EARTH 150 (Fall 2007) and GEOSC 402y (Natural Disasters; Spring 2008, Fall 2009), PSU

GRANTS & AWARDS
Peter Deines Lectureship, First prize talk by a Post-Comps PhD student 2011
Chesapeake energy scholarship 2011
Shell research award 2009, 2010
SUNY Brockport Student Writing Award for Scientific and Technical Writing 2007
Eagle Scout Award 2003

PROFESSIONAL & SERVICE ACTIVITIES
• Co-chair of the Department of Geosciences Colloquium, Penn State, 2011-2012
• Reviewer for the International Journal of Solids and Structures
• Assistant Scoutmaster, Juniata Valley Council, Troop 367, 2008-2012

PUBLISHED PAPERS
• Haines, SH, BM Kaproth, C Marone, D Saffer, and B van der Pluijm, Shear zones in clay-rich fault gouge: A laboratory study of fabric element development and evolution. JSG (Accepted, 2013)
• Johnson, PA, B Carpenter, M Knuth, BM Kaproth, P-Y Le Bas, R. Guyer, EG Daub, and C Marone [2012], Nonlinear dynamical triggering of slow-slip on simulated earthquake faults with implications to Earth, JGR
• Kaproth, BM, S. Cashman, and C. Marone [2010], Deformation band formation and strength evolution in un lithified sand: the role of grain breakage, JGR

RECENT TALKS
• Kaproth-Gerecht, BM [2013], Very slow lab earthquakes with precursor sound speed changes, SSA #13-712
• Deines Lecture, invited seminar, Department of Geosciences, Penn State [2012] Deformation Bands: How they mess with oil companies, and three ways to prepare for them
• Kaproth, BM, and C Marone [2011], Fault gouge evolution during rupture and healing: Continual active-seismic observations across laboratory-scale fault zones, 2011 AGU, #S23D-04
• Kaproth, BM, C Marone [2010], Quantifying pressure solution and lithification: tying elastic moduli measurements to changes in porosity and deformation style in sheared granular aggregates, AGU, #T32A-04