The Pennsylvania State University
The Graduate School
College of Earth and Mineral Sciences

AN EXPERIMENTAL INVESTIGATION OF FRICTIONAL AND HYDRAULIC PROPERTIES OF SHEAR ZONES, WITH APPLICATION TO EARTHQUAKE FAULTS AND GLACIAL TILL

A Dissertation in
Geosciences

by

Andrew Paul Rathbun

© 2010 Andrew P. Rathbun

Submitted in Partial Fulfillment of the Requirements for the Degree of Doctor of Philosophy

December 2010
The dissertation of Andrew P. Rathbun was reviewed and approved* by the following:

Chris J. Marone  
Professor of Geosciences  
Associate Head for Graduate Programs and Research  
Dissertation Adviser

Sridhar Anandakrishnan  
Professor of Geosciences

Richard B. Alley  
Evan Pugh Professor of Geosciences

Derek Elsworth  
Professor of Energy and Geo-Environmental Engineering

Demian Saffer  
Associate Professor of Geosciences

*Signatures are on file in the Graduate School.
Faulting in the brittle crust is controlled by rate and state friction (RSF) and fluid migration. In a series of four manuscripts, this work explores shear zones with a series of laboratory experiments on both natural and synthetic materials. I conduct shear experiments to investigate how localization of shear controls the gouge zone of faults. Experiments show that shear localizes progressively into a central zone. Localized shear can lead to a change in the RSF parameters where micromechanics of the shear zone overrides the expectations of the commonly used laws. I also find that subtle changes in the fabric of the shear zone can enhance the possibility of seismic slip. I find that slow, seismic slip can be produced in laboratory experiments as creep rupture rather than strictly RSF and stick-slip sliding. Acoustic emissions of slow-slip have a similar form to stick-slip; however, the duration is ~1s rather than ~1ks observed in laboratory stick-slip. In my experiments it is possible to propagate slow-slip in both velocity-strengthening and weakening materials. Elevated fluid pressure in gouge zones can mitigate the effects of the frictional behavior by increasing pressure and thus decreasing effective stress though thermal pressurization or by decreasing fluid pressure by dilatancy hardening. Tests on fault gouge from the San Andreas shows that the fault core has low permeability. The San Andreas would act as a barrier to fluid flow and could behaved as an undrained zone leading to dilantant hardening or thermal weakening. The results of this dissertation are an important step in understanding fault behavior from stable (aseismic) sliding to slow-slip and finally stick-slip (seismic) sliding.
# TABLE OF CONTENTS

List of Figures ........................................................................................................ vii
List of Tables ........................................................................................................ viii
Acknowledgements ............................................................................................... ix

Chapter 1. INTRODUCTION ............................................................................. 1
  1.1. INTRODUCTION ................................................................................... 1
  1.2. BACKGROUND ...................................................................................... 2
  1.3. SUMMARY OF CHAPTERS ................................................................... 6
References .............................................................................................................. 7

Chapter 2. EFFECT OF STRAIN LOCALIZATION ON FRICTIONAL BEHAVIOR
OF SHEARED GRANULAR MATERIALS ............................................................. 9
ABSTRACT .......................................................................................................... 10
  2.1. INTRODUCTION ................................................................................... 11
  2.2. EXPERIMENTAL METHODS ................................................................ 14
    2.2.1 Procedure for monitoring strain localization .................................... 17
  2.3. PROCEDURE, RESULTS AND ANALYSIS OF EXPERIMENTS ....... 18
    2.3.1 Creep experiments ........................................................................ 18
    2.3.2 Dilation and the onset of localization .............................................. 20
    2.3.3 Slip velocity step tests ...................................................................... 21
    2.3.4 Evolution of the critical slip distance .............................................. 23
    2.3.5 Strain markers and localized deformation ...................................... 26
  2.4. DISCUSSION .......................................................................................... 29
    2.4.1 Dilation as a proxy for shear localization ....................................... 29
    2.4.2 Symmetry of frictional behavior for velocity increases and decreases 33
    2.4.3 Localization and till rheology ......................................................... 36
  2.5. CONCLUSIONS ...................................................................................... 37
  2.6. ACKNOWLEDGMENTS ....................................................................... 38
REFERENCES ...................................................................................................... 38

Chapter 3. SYMMETRY IN RATE AND STATE FRICTION ................................ 58
ABSTRACT .......................................................................................................... 59
  3.1. INTRODUCTION ................................................................................... 60
    3.1.1 Comparison of Evolution Laws ....................................................... 63
Appendix A: Strain localization in granular fault zones at laboratory and tectonic scales

Appendix B: Earthquake energy budget

Appendix C: Matlab codes
   1. Stick-slip picker
   2. Creep rate calculator
# List of Figures

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Figure 2.1</td>
<td>Double direct shear configuration</td>
<td>45</td>
</tr>
<tr>
<td>Figure 2.2</td>
<td>Grain size distribution</td>
<td>46</td>
</tr>
<tr>
<td>Figure 2.3</td>
<td>Creep experiment</td>
<td>47</td>
</tr>
<tr>
<td>Figure 2.4</td>
<td>Layer and strain response to a stress step</td>
<td>48</td>
</tr>
<tr>
<td>Figure 2.5</td>
<td>Dilation as a function of step size</td>
<td>49</td>
</tr>
<tr>
<td>Figure 2.6</td>
<td>Dilation as a function of strain</td>
<td>50</td>
</tr>
<tr>
<td>Figure 2.7</td>
<td>Velocity stepping experiment</td>
<td>51</td>
</tr>
<tr>
<td>Figure 2.8</td>
<td>Sensitivity analysis of Dc</td>
<td>52</td>
</tr>
<tr>
<td>Figure 2.9</td>
<td>Rate and state parameters of till</td>
<td>53</td>
</tr>
<tr>
<td>Figure 2.10</td>
<td>Shear localization in a granular experiment</td>
<td>54</td>
</tr>
<tr>
<td>Figure 2.11</td>
<td>Model of localization</td>
<td>55</td>
</tr>
<tr>
<td>Figure 2.12</td>
<td>Dilation as a proxy for localization</td>
<td>56</td>
</tr>
<tr>
<td>Figure 2.13</td>
<td>Asymmetry in velocity steps</td>
<td>57</td>
</tr>
<tr>
<td>Figure 3.1</td>
<td>Expected frictional response to a velocity step</td>
<td>87</td>
</tr>
<tr>
<td>Figure 3.2</td>
<td>Grain size distribution</td>
<td>88</td>
</tr>
<tr>
<td>Figure 3.3</td>
<td>Velocity stepping experiment and rate and state parameters</td>
<td>89</td>
</tr>
<tr>
<td>Figure 3.4</td>
<td>Factor of 3 velocity steps</td>
<td>90</td>
</tr>
<tr>
<td>Figure 3.5</td>
<td>Factor of 30 velocity steps</td>
<td>91</td>
</tr>
<tr>
<td>Figure 3.6</td>
<td>Dilation for factor of 3 velocity steps</td>
<td>92</td>
</tr>
<tr>
<td>Figure 3.7</td>
<td>Dilation for factor of 30 velocity steps</td>
<td>93</td>
</tr>
<tr>
<td>Figure 3.8</td>
<td>Bare surface velocity steps</td>
<td>94</td>
</tr>
<tr>
<td>Figure 3.9</td>
<td>Stress oscillation experiment</td>
<td>95</td>
</tr>
<tr>
<td>Figure 3.10</td>
<td>Velocity steps before and after oscillations</td>
<td>96</td>
</tr>
<tr>
<td>Figure 4.1</td>
<td>Characteristic duration of seismic events</td>
<td>108</td>
</tr>
<tr>
<td>Figure 4.2</td>
<td>Laboratory stick-slip</td>
<td>109</td>
</tr>
<tr>
<td>Figure 4.3</td>
<td>Slow-slip events vs. rate and state friction</td>
<td>110</td>
</tr>
<tr>
<td>Figure 4.4</td>
<td>Seismograms of laboratory slip events</td>
<td>111</td>
</tr>
<tr>
<td>Figure 4.5</td>
<td>Stress drop as a control of velocity and slip</td>
<td>112</td>
</tr>
<tr>
<td>Figure 4.81</td>
<td>Grain size distribution</td>
<td>113</td>
</tr>
<tr>
<td>Figure 4.82</td>
<td>Stick-slip and slow-slip in experiment p2773</td>
<td>114</td>
</tr>
<tr>
<td>Figure 5.1</td>
<td>Permeability apparatus</td>
<td>134</td>
</tr>
<tr>
<td>Figure 5.2</td>
<td>Pulse decay method</td>
<td>136</td>
</tr>
<tr>
<td>Figure 5.3</td>
<td>Constant rate of strain experiment</td>
<td>137</td>
</tr>
<tr>
<td>Figure 5.4</td>
<td>Flow-through test</td>
<td>138</td>
</tr>
<tr>
<td>Figure 5.5</td>
<td>Berea Sandstone permeability</td>
<td>139</td>
</tr>
<tr>
<td>Figure 5.6</td>
<td>Crab Orchard Sandstone permeability</td>
<td>140</td>
</tr>
<tr>
<td>Figure 5.7</td>
<td>Wilkeson Sandstone permeability</td>
<td>141</td>
</tr>
<tr>
<td>Figure 5.8</td>
<td>Geologic setting</td>
<td>142</td>
</tr>
<tr>
<td>Figure 5.9</td>
<td>Permeability of the San Andreas Fault</td>
<td>143</td>
</tr>
<tr>
<td>Figure 5.10</td>
<td>Specific storage of fault gouge</td>
<td>144</td>
</tr>
<tr>
<td>Figure 5.11</td>
<td>Young’s modulus</td>
<td>145</td>
</tr>
</tbody>
</table>
## List of Tables

<table>
<thead>
<tr>
<th>Table</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Table 2.1</td>
<td>Experiment table</td>
<td>44</td>
</tr>
<tr>
<td>Table 3.1</td>
<td>Experiment table</td>
<td>86</td>
</tr>
<tr>
<td>Table 4.1</td>
<td>Experiment table</td>
<td>107</td>
</tr>
<tr>
<td>Table 5.1</td>
<td>Interlab comparison</td>
<td>132</td>
</tr>
<tr>
<td>Table 5.2</td>
<td>Experiment table</td>
<td>133</td>
</tr>
</tbody>
</table>
Acknowledgments

I would like to thank all of the great people I have gotten to work with in the lab over the years, and they are too numerous to list here. All of their help and patience both helped advance this dissertation and make my time at Penn State enjoyable. In particular I want to thank Jon Samuelson who always was a willing listener to all of my crazy ideas (and many complaints) and the times I couldn’t quite articulate my thoughts. Doug Edmonds provided a lot of valuable insight both to my dissertation and on several other things. My parents and family were very understanding along the way, even if they never quite understood what I do or why I do it. It was all those trips around the country wondering how the hills got there and what that funny rock was in Canada with the hard parts sticking out of it that inspired me to try geology. Without my family, I never would have been able to get here.

My committee has been great. Richard Alley, Sridhar Anandakrishnan and Derek Elsworth were always ready and willing to help me at any time. Demian Saffer put in a lot more work than just a committee member and was really a co-advisor during my last couple years. Chris Marone has really been a great advisor and a lot more. His willingness to let me go off on my own and fiddle in the lab until I figured something out was great. I can’t imagine someone that could have been more patient with my ever-shifting ideas and promises of things to do for this dissertation.
Chapter 1: INTRODUCTION

1.1 INTRODUCTION

Brittle shear zones control two of the most dynamic systems in the earth, glacial motion and earthquakes. Mature fault zones contain a gouge zone of crushed and altered material ranging from microns to meters thick that hosts earthquakes. Some glacial systems contain a layer of till underneath the ice, which can deform and lead to rapid motion. While these two environments seem different, the same underlying physics seems to control them. This thesis aims to understand some of those fundamental controls in brittle media whether it is the gouge zone of a fault system or the deformable till beneath a glacier.

This thesis is written as a series of manuscripts. Each manuscript is a laboratory treatment of brittle shear aimed at understanding the fundamental controls of gouge and till zones. This chapter is a very brief background of friction and strength in shear zones. I provide the framework for our understanding of how things slide and my main interest, how earthquakes begin. I do not attempt to provide a comprehensive review of friction, but merely to show some background of the tools I use. Chapter 2 deals with how shear localizes into a narrow zone and introduces the concept of symmetry and asymmetry in friction. Chapter 3 takes an in-depth look at Rate and State Friction (RSF) in the lab, attempting to unravel how friction evolves. Chapter 4 deals with the differences between stick-slip and slow-slip events in the lab and proposes a new mechanism for studying slow-slip. Chapter 5 looks at the permeability of the San Andreas Fault by measuring across fault permeability which
coupled with RSF is typically invoked as a mechanism for slow-slip. Appendix A presents a paper in which I am the secondary author on localization at the lab and fault scale, Appendix B is another paper in which I am a secondary author. It deals with the energy partitioning in earthquakes. Appendix C is reference codes for analyzing data.

1.2 BACKGROUND

Our understanding of earthquake propagation in brittle shear zones is based on the laws of friction. Gillaume Amontons put forth the first two laws: 1) Frictional force is directly proportional to an applied normal load and 2) Frictional force is independent of the area of contact. More simply this is expressed as

\[ F_{\text{fric}} = \mu F_n \]  

(1).

Coulomb extended this idea to include that the frictional force is independent of rate. Coulomb also explained a larger static friction that kinetic friction by the interlocking of asperities, the roughness of the shear contact. The assertion of Coulomb that friction is rate independent is true until you look at fine-scale measurements in ~ the third decimal place, which forms the basis modern rate and state friction.

Rabinowitz [1958] looked at the evolution of from static to dynamic friction and formed the basis for modern rate and state friction. The evolution was based on asperities and contact size. When two surfaces are in contact with each other, the strength of that contact grows due to an increase in the size of the real area of contact, \( A_r \).
For sliding bodies, two styles of motion are possible 1) steady, stable sliding in which motion is constant and occurs at a steady rate (aseismic) and 2) stick-slip in which the bodies are locked together and then suddenly slip past one another (seismic). We deal with these two situations on daily basis. You can slide your hand over a smooth table and it is likely to slide stability, if you push down harder it is likely to start to stick-slip, the sound of a bow on a violin string, or the screeching of fingernails on a chalkboard are the result of stick-slip.

We describe which style of sliding is likely to occur with the second-order frictional variations described by the RSF equations [Dieterich, 1978; 1981; Ruina 1983]. RSF forms our basis for understanding earthquake generation. In RSF, friction is a function of velocity and a state variable such that:

\[
\mu = \mu(V,\theta) = \mu_0 + a \ln\left(\frac{V}{V_0}\right) + b \ln\left(\frac{\theta}{D_c}\right)
\]

(2)

where \(V\) is velocity, \(\theta\) is a state variable, \(\mu_0\) is the background sliding friction at velocity \(V_0\). The direct effect, \(a\), describes the increase (decrease) in friction after a velocity increase (decrease), the evolution, \(b\), is the decrease (increase) of friction from the peak (trough) after the velocity change. The \(e\)-folding distance to reach the new steady state from \(a\) to \(b\) is the critical slip distance, \(D_c\). The time evolution of the state variable is usually described using one of two common laws:

\[
\frac{d\theta}{dt} = 1 - \frac{V\theta}{D_c}
\]

(3)

\[
\frac{d\theta}{dt} = -\frac{V\theta}{D_c} \ln\left(\frac{V\theta}{D_c}\right)
\]

(4)
At steady state sliding, the combination of (2) and either (3) or (4) gives the velocity dependence of friction \((a-b)\)

\[
a - b = \frac{\Delta \mu_{se}}{\Delta \ln V}.
\]

When \((a-b)\) is positive the material is velocity-strengthening and when \((a-b)\) is negative the material is velocity-weakening.

Velocity-strengthening, weakening and the possibility of stick-slip can be understood intuitively. If a body (in our case a fault or glacier) starts to slide faster and friction, the resistance to motion, increases, it will inhibit motion and serve to arrest the bodies. In the other case, when sliding velocity is increased and the frictional strength of the contact decreases an earthquake is possible.

Velocity-weakening can be attributed to a decrease in \(A_r\) [Rabinowitz, 1958]. Faster motion decreases the contact time, and as a result the frictional contact does not grow to as large of a value. The alternative situation when with increased velocity friction increases another mechanism must control friction. Typically, dilantancy [e.g. Mead 1925; Frank, 1965] is invoked to explain extra work needed after a change in velocity to explain strengthening.

The propagation of an earthquake is possible when the critical stiffness, \(k_c\), exceeds the background level, \(k\). Critical stiffness is defined by the rate and state terms as

\[
k < k_c = \frac{-(a-b)\sigma_n}{D_c}.
\]

This is an analogous situation to a simple spring and slider. If the spring is sufficiently compliant, it stretches before the slip occurs until the pull of the spring
exceeds the strength of the contact. Once the stored energy is released the slider stops. If the spring is stiff the force is transferred to the frictional contact, the amount of energy stored is smaller and the slider continuously displaces because the pull of the spring always exceeds the strength of the slider on a surface. Critical stiffness brings in not only \((a-b)\) but also the normal stress \(\sigma_n\) and \(D_c\). Understanding the influence of normal stress is intuitive if you once again think about a hand sliding on table. If you barely push down your hand is likely to slide stably. If you push harder, and it is a dry day so no moisture is lubricating the contact, your hand will probably stick-slip. This process extends to the piece of chalk, a pencil eraser, or many frictional contacts that we use every day.

It is important to remember that few places in the brittle crust or base of a glacier exist in the absence of fluids. The presence of fluid can, not only control the RSF parameters \((a-b)\) and \(D_c\) be interacting with the minerals, but also influences the stresses. Fluids and pore pressure reduce the stress the stress felt by the surrounding material. Dilation and compaction of a fault zone can decrease and increase the pore pressure of a shear zone, respectively. The Coulomb-Mohr relation governs brittle failure.

\[
\tau_f = \mu(\sigma_n - Pp) + c
\]  

(7)

Where \(\tau_f\) is a shear strength \(Pp\) is pore pressure and \(c\) is cohesion. A shear zone with sufficiently low permeability can behave in the undrained state. If dilation occurs (increased porosity or void volume) the \(Pp\) can decrease due to a greater volume for the fluid to fill. This can serve to arrest fault motion and is one of the mechanisms invoked to explain slow-slip.
1.3 SUMMARY OF CHAPTERS

The aim of this dissertation is to understand the frictional deformation of shear zones. Chapter 2 looks at localization in granular media. I explore the evolution of localization with shear. In granular shear zones containing many particles and interactions between them, localization reduces shear in to a narrow zone or even a plane. The grain-to-grain velocity-weakening interactions control the system rather than dilation and many grains leading to velocity-strengthening. I show that localization can occur in a velocity-strengthening material that dilates. Localization occurs progressively into a boundary parallel zone a few grain diameters thick.

Chapter 3 expands on some observations from Chapter 2 in that the frictional response to a velocity step is asymmetric. I find that the Ruina law (Equation 4) better fits laboratory data; however variations exist that are not predicted by either of the two evolution laws. It is also shown that oscillations and variations in the shearing layer can decrease the critical slip distance, which increases the likelihood of an earthquake (i.e. Equation 6).

Chapter 4 explores the transition from seismic stick-slip to transient slow-slip events in the lab. Constant shear stress experiments produce transient events in both velocity-strengthening and weakening materials, an unexpected result from RSF (i.e. Equation 6). I do find that the slip duration scales with \((a-b)\) with positive \((a-b)\) producing the longest slip duration (100’s seconds) and the most negative \((a-b)\) producing the shortest (seconds). I propose a new mechanism for slow-slip in which
creep failure and rupture can lead to transient events rather than purely the interplay of RSF and high fluid pressure.

The final chapter is unique in this dissertation in that permeability experiments rather than shear experiments are presented. Experiments were conducted in multiple configurations to explore faults as fluid barriers or pathways. These experiments represent the first permeability measurements from the gouge zone of the San Andreas Fault Observatory at Depth (SAFOD) drilling project. These experiments help constrain the possibility that fluid pressurization can induce seismicity on in the fault or that depressurization due to dilation can stabilize the fault.

Appendices A and B are two papers in which I am the second author. Both present shear experiments and model results trying to understand the physics of earthquakes. These two papers relate to all four main chapters of my dissertation and present new treatments of experimental fault zones.

REFERENCES


Mead, W. J. (1925), The geologic role of dilatancy, *J. Geol.* 33, 685-698.


Chapter 2: EFFECT OF STRAIN LOCALIZATION ON FRICIONAL BEHAVIOR OF SHEARED GRANULAR MATERIALS

Andrew P. Rathbun and Chris Marone

Rock and Sediment Mechanics Laboratory, Penn State Center for Ice and Climate, The Pennsylvania State University, University Park, PA 16802, USA

Submitted to the Journal of Geophysical Research, 19 March 2009
Resubmitted to the Journal of Geophysical Research, 24 September 2009
Published in the Journal of Geophysical Research, January 2010
ABSTRACT: We performed laboratory experiments to investigate shear localization and the evolution of frictional behavior as a function of shear strain. Experiments were conducted on water-saturated layers, 0.3 to 1 cm thick, of Caesar till, a granular material analogous to fault gouge. We used the double direct shear configuration at normal stresses ranging from 0.5-5 MPa and shearing velocities of 10-100 μm/s. Shear localization was assessed via strain markers and two proxies: 1) macroscopic layer dilation in response to perturbations in shear stress and 2) rate/state friction response to shear velocity perturbations. In creep-mode experiments, at constant shear stress, we monitored dilation for perturbations in shear stress. In standard friction tests, we measured the coefficient of friction during perturbations in macroscopic strain rate. We find evidence of strain localization beginning at shear strain $\gamma \approx 0.15$ and continuing until $\gamma \approx 1$. Analysis of strain markers support interpretations based on the proxies for localization and show that strain is localized in zones of finite thickness. We also investigate symmetry of the friction response to step changes in imposed slip velocity and find that the behavior is symmetric. Our results, favor the Ruina law for friction state evolution, in which slip is the fundamental variable, rather than the Dieterich law. The critical slip distance for friction evolution, $D_c$, is $\sim 140$ μm. The Dieterich state evolution law requires different values of $D_c$ for velocity increases/decreases, 100 μm vs. 175 μm, respectively, and would imply strain localization/delocalization associated with shear in a finite zone.
2.1 INTRODUCTION

The localization of strain in brittle shear zones has broad implications for the seismic and aseismic nature of tectonic faulting and the rheology of subglacial till. Cataclasitic processes, wear, and reworking of sediments form fault gouge and its analog in the deforming substrate of some glaciers, subglacial till. Field observation of brittle shear zones [e.g. Logan et al., 1979; Arboleya and Engelder, 1995; Chester and Chester, 1998; Cashman and Cashman, 2000; Faulkner et al., 2003; Hayman et al., 2004; Fossen et al., 2007; Cashman et al., 2007] laboratory experiments [e.g. Mandl et al., 1977; Logan et al., 1979; 1992; Marone et al., 1990; Gu and Wong, 1994; Beeler et al., 1996; Scruggs and Tullis, 1998; Niemeijer and Spiers, 2005], and numerical simulations [e.g. Antonellini and Pollard, 1995; Morgan and Boettcher, 1999; Mair and Abe, 2008] show that slip often localizes into discrete zones or along distinct fabrics in the shear zone.

Field observations from exhumed brittle shear zones indicate that large slip (10’s of km) may occur in zones that range in width from a few centimeters to 10’s of meters [Mooney and Ginzburg, 1986; Montgomery and Jones, 1992; Chester and Chester, 1998; Storti et al., 2003; Sibson, 2003; Wibberley and Shimamoto, 2003; Billi and Storti, 2004; Chester et al., 2004; Di Toro et al., 2005]. Fault zones grow in width by continued slip and evolve internally due to grain size reduction and mineral growth along shear bands [e.g. Engelder, 1974; Scholz, 1987; Schleicher et al., 2006]. Such faults often include highly localized principle slip zones, which are hosted in fault damage regions and gouge zones. Fault zone width is difficult to quantify and
exhibits extreme variation along strike, even for a single fault, but generally ranges from centimeters to 100’s of meters or more [e.g. Scholz, 2002; Sibson, 2003].

Rate and state friction has been used to describe the behavior of brittle faulting in gouge and rocks [Dieterich, 1979; 1981; Ruina, 1983] based on the idea that stick-slip motion and interseismic creep is an analog for the seismic cycle [Brace and Byerlee, 1966]. Frictional instability requires that faults weaken with either increased slip (slip weakening) or increased velocity (velocity weakening). If the rate of weakening exceeds a critical value, elastic strain energy is released from the surrounding materials, causing shear heating and elastic wave radiation. The critical weakening rate depends on the normal stress and elastic properties of the fault region [e.g. Scholz, 2002]. For deformation zones that exhibit increased frictional strength with increasing strain rate (so called velocity strengthening frictional behavior) aseismic slip and creep are expected. Such behavior is expected for pervasive deformation prior to strain localization [Marone et al., 1990; 1992]. The term creep is often used in two different contexts; in this study we will use the word creep to denote deformation under constant shear stress, rather than to distinguish aseismic from seismic slip. In a granular material, pervasive shear and velocity strengthening frictional behavior have been attributed to the dilational work required to expand the layer [Mead, 1925; Frank, 1965; Marone et al., 1990].

Many of the processes that govern friction and strain localization in fault gouge also appear to be important in subglacial till. Shear deformation within till accounts for a large portion of the net displacement of some fast moving glaciers and ice streams [e.g. Clarke, 2005]. The rheology of subglacial till has been a matter of
much debate; see [Alley, 2000] and [Clarke, 2005] for recent reviews. Early investigators used a power law relationship for glacial till where strain rate is proportional to shear stress raised to a stress exponent, \( n \) [e.g. Boulton and Hindmarsh, 1987]. For convenience most modeling studies have assumed that till behaves as a viscous material with \( n \) of order 1, whereas most laboratory studies report a frictional (often termed ‘plastic’) rheology of \( n > 15 \) [e.g. Kamb, 1991; Iverson et al., 1997; 1998]. Work by Rathbun et al. [2008] shows that the rheology of till evolves from \( n < 10 \) to \( n > 50 \) from the onset of motion to steady frictional sliding. Recent studies show that glaciers exhibit stick-slip motion in some cases [Anandakrishnan and Bentley, 1993; Ekstrom et al., 2003; 2006] and physical models have been proposed [Tsai et al. 2008; Winberry et al., 2009]. Basal tills are often characterized by zones of localized shear, [e.g. Truffer et al., 2000; Kamb 2001; Evans et al., 2006; Menzies et al., 2006] and laboratory studies indicate that till rheology is governed in part by the distribution of strain localization [Larsen et al., 2006; Thomason and Iverson, 2006; Iverson et al., 2008; Rathbun et al., 2008]. However, there are relatively few detailed laboratory studies of strain localization and its effect on friction constitutive properties of till.

Laboratory studies focused on earthquake faulting have shown that fault gouge often exhibits a transition from pervasive to localized deformation with increasing strain and that this transition coincides with a change from stable to stick-slip frictional sliding [e.g. Logan et al., 1979; 1992; Marone et al., 1990; Beeler et al., 1996; Marone, 1998]. Similar connections between strain localization and frictional properties are emerging from studies of glacial till [e.g. Iverson et al., 2008;
However, most studies of till do not include direct information on friction constitutive behavior or stick-slip.

The purpose of this paper is to report on laboratory investigations of strain localization and its influence on frictional behavior of till and granular fault gouge. We employ both constant shear velocity and constant shear stress boundary conditions, with careful attention to the influence of shear fabric development on frictional strength, layer dilation, and rate/state friction properties including the critical slip distance and steady-state frictional strength.

2.2 EXPERIMENTAL METHODS

Experiments were performed in a servohydraulic, biaxial testing apparatus using the double-direct shear configuration (Figure 2.1). Two granular layers were sheared simultaneously between three steel forcing blocks at constant normal stresses of 0.5, 1, and 5 MPa (Table 2.1). The horizontal ram of the testing machine applies a constant normal force and the vertical ram imposes shear traction. Both rams can operate in stress or displacement servocontrol. Layer dimensions were 10 cm x 10 cm (nominal frictional contact area) x a thickness of 0.3, 0.5 or 1 cm (Table 2.1). Forcing blocks were grooved to a depth of 0.8 mm with wavelength of 1 mm perpendicular to shear to ensure that deformation occurred within the layer and not at the layer-block interface.

Granular layers were constructed by applying a wall of cellophane tape around the forcing blocks to the desired layer thickness. The sample was then added and planed off to the desired thickness using a precision leveling jig (Table 2.1). This
method produced constant initial layer thickness to a tolerance of ±0.2 mm. To reduce material loss along the front/back layer edges, guide plates were attached to the side forcing blocks. Molybdenum lubricant was used beneath the side forcing blocks to facilitate motion and layer dilation/compaction at constant normal stress. To further reduce material loss, a 0.01” latex sheet was applied to the underside of the blocks. Calibration experiments show that the latex sheet adds < 20 N (2 kPa) to the measured shear force [Carpenter, 2007] and we correct for this effect along with the gravitational force associated with the mass of the center-forcing block (19 N).

Strain markers were constructed in select experiments (Table 2.1) by placing 3 sets of brass sheets (0.005” thick) at equally spaced increments in the layer. Each set was filled with a 2-mm wide layer of blue sand (Kelly’s Crafts Inc.) and then the brass sheets were removed leaving a strip blue sand in the layer. The bulk weight percentage of markers was kept <5% to ensure that this material did not impact bulk frictional strength of the layer [e.g. Logan and Rauenzahn, 1987].

All experiments were conducted under nominally saturated conditions by submerging the sample in water using a flexible rubber membrane (Figure 2.1). The reservoir was left open to the atmosphere at the top, resulting in saturated drained-conditions for the layer. Before the application of stress, the sample was allowed to equilibrate with water for at least 45 minutes to ensure complete saturation.

Normal and shear forces were measured with BeCu load cells to 0.1kN resolution. Displacements were measured by direct current displacement transducers (DCDT’s) to 0.1 µm resolution. Experiments were recorded with 24-bit analog-to-digital precision. We used a recording rate of 10 kHz and averaged samples for
storage at > 10 sample per micron of shear displacement in all experiments. Shearing velocity was computer controlled via an analog servo-command signal updated at 100 Hz. The initial layer thickness was measured, in situ under load, to ±0.01 mm. Measured displacements normal to the layer correspond to changes in layer thickness at constant normal stress. Both normal and shear displacements reported here have been corrected for the elastic stiffness of the vertical and horizontal load frames, 5 MN/cm and 3.7 MN/cm, respectively. We measure macroscopic shear strain of the layer by integrating the measured slip increments, imposed at the layer boundaries, and dividing by the instantaneous layer thickness.

\[
\gamma = \frac{X_{\text{max}}}{h_i},
\]

where \( \gamma \) is bulk shear strain, \( x_i \) is the position of the center forcing block (e.g. shear offset at the layer edge), \( h \) is the instantaneous thickness at slip increment \( i \), and \( X_{\text{max}} \) is the total displacement.

The experiments were conducted using Caesar till, which is a mixed grain-size granular material that derives from the Scioto (Ohio) Lobe of the Laurentide Ice Sheet and dates to ~19,500 years ago [Haefner, 2000]. Samples were air-dried and then disaggregated by hand before grain-size analysis following the procedures of Rathbun et al. [2008]. We sieved the till and removed the >1mm fraction, in order to preserve stress homogeneity at our layer boundaries and to ensure that deformation was representative of the sample as a whole, rather than a few large grains. The experimental grain size was 98.7% sand 1.2% silt and 0.1% clay-sized grains (Figure 2.2) with some grains composed of aggregations of several small particles. Grain sizes less than 63 \( \mu \)m were analyzed using laser obscuration in a Malvern
Mastersizer. Sample composition was determined via X-ray diffraction \cite{Underwood et al., 2003}, with relative abundances of 35\% quartz, 26\% calcite 23\% plagioclase, and 16\% clay minerals with the clay particles composed of 35.3\% smectite, 38.5\% illite and 26.1\% chlorite/kaolinite.

### 2.2.1 Procedure for monitoring strain localization

Strain localization and shear fabric evolution were investigated by direct observation of preserved samples, post shear, and by indirect metrics measured in-situ during the experiment. Layers were impregnated with epoxy for microstructural evaluation and tracking of strain markers. Experiments at low normal stresses and with granular particles typically do not show a well-developed shear fabric \cite[e.g. Mair and Marone, 1999]{Mair and Marone, 1999}. Therefore, we developed indirect methods of fabric characterization based on the layer response to perturbations in shearing rate and shear stress. These include layer dilation, friction memory effects characterized by the critical slip distance, and the steady-state rate dependence of kinetic friction.

Layer dilation was used as a proxy for strain localization. That is, we assume that only the fraction of the layer undergoing active strain exhibits shear dilation upon a perturbation in loading rate. Pervasive shear, in which the whole layer is actively involved in shear, results in larger dilation than localized shear, in which only a fraction of the layer is actively involved in shear. We measure dilation after accounting for geometric thinning of the layer in direct shear \cite[Scott et al., 1994]{Scott et al., 1994}. 
2.3 PROCEDURE, RESULTS AND ANALYSIS OF EXPERIMENTS

We conducted two types of experiments for this study (Figure 2.3): 1) creep mode tests in which layers were sheared at a controlled shear stress value, and 2) standard friction tests in which layer where sheared at a controlled shear displacement rate.

2.3.1 Creep experiments

Creep mode shearing (Figure 2.3A) began by first measuring steady state frictional strength, \( \tau_{res} \) (Figure 2.3B) at constant shear displacement rate. We refer to the shear strain that accumulated prior to creep loading as preconditioning shear strain, \( \gamma_i \) and we varied \( \gamma_i \) from 0 to 4.3 to investigate its effect on fabric development and creep rheology. Creep tests began at a shear stress of \( \sim 70\% \) of \( \tau_{res} \) and shear stress was increased in steps equal to 2\%, 5\% or 7.5\% of \( \tau_{res} \) and held for 45 minutes (Table 2.1). For experiments that did not reach steady shear strength during \( \gamma_i \), \( \tau_{res} \) was estimated using an average value from other experiments [Rathbun et al., 2008] and then checked after creep loading.

We measured frictional rheology and layer thicknesses changes at each stress until tertiary creep occurred (Figure 2.3). For shear stresses < 90\% of \( \tau_{res} \), shear strain was negligible during creep step tests [Rathbun et al., 2008]. However, tertiary creep produced measurable shear strain for stresses near \( \tau_{res} \), as shown for the final stress step in Figure 2.3; note that \( \sim 18 \text{ ks} \) in Figure 2.3A corresponds to \( \gamma \sim 1.2 \) in Figure 2.3B. For the conditions of our study, tertiary creep began at 92-100\% of \( \tau_{res} \), depending on \( \gamma_i \) [Rathbun et al., 2008].
After completion of the creep portion of the experiment, layers were again sheared at a constant displacement rate of 10 μm/s to investigate strain hardening and changes in friction (Figure 2.3A). The difference in frictional strength before and after creep was always <1.5%, and thus we assume that the strain accumulation during creep tests did not significantly affect fabric development.

Details of the creep behavior during stress steps are given in Figure 2.4. The stress step rise time was 1-2 sec, during which time shear strain rate increased rapidly. Layer dilation is clearly evident in the raw data dashed line in (Figure 2.4), but to improve measurement precision we removed the trend in layer thickness associated with geometric thinning [Scott et al., 1994]. For a step increase in stress, strain rate first increased, consistent with primary creep, and then decayed steadily to a steady-state value within 30 to 40 minutes (Figure 2.4 inset), which we associated with secondary creep. We did not attempt to verify the establishment of secondary creep in each case, because many previous works have shown that friction of geomaterials exhibits log-time creep relaxation for subcritical stresses [e.g. Marone, 1998; Karner and Marone, 2001; Mitchell and Soga, 2005; Rathbun et al., 2008]. However, for the resolution of our measurements (< 0.1 μm) layer dilation was complete within 10 to 20 minutes after a stress step (Figure 2.4 inset). We define creep dilation Δh* as the difference between layer thickness 40 minutes after the stress step and initial layer thickness before the step. Positive Δh* represents dilation. The values of Δh* do not vary systematically as a function of shear stress in a given experiment.
2.3.2 Dilation and the onset of localization

We investigated the effect of shear localization on creep behavior by systematically varying preconditioning shear strain $\gamma_i$ (Figure 2.3). Figure 2.5 shows data from 12 experiments in which we compare creep dilation as a function of stress step magnitude and shear strain. Our three stress step magnitudes range from 0.01 to 0.045 MPa (Figure 2.5). Layer dilation $\Delta h^*$ increased strongly with stress change for layers with low initial strain ($\gamma_i < 0.2$), whereas for higher values of $\gamma_i$ dilation was nearly independent of stress step size (Figure 2.5).

To further investigate shear localization and fabric development, we analyzed the effect of stress perturbations on layer thickness $\Delta h^*$ (e.g. Figure 2.4) as a function of $\gamma_i$ (Figure 2.6). The dilation parameter $\Delta h^*$ is a proxy for fabric development if dilation occurs within only the fraction of the layer thickness where strain is localized. Figure 2.6 shows data for three layer thicknesses and two normal stresses. For our range of conditions $\Delta h^*$ did not vary systematically with normal stress (Table 2.1). Each point in Figure 2.6 represents the average of all shear stress steps in a given experiment plotted vs. $\gamma_i$ (e.g. Figure 2.3).

The bulk of our experiments were conducted using 1-cm thick layers. In these experiments, creep dilation decreased systematically as a function of initial shear strain and reached stable values by $\gamma_i \sim 1.2$-2 (Figure 2.6). Layer dilation was about 6 $\mu$m for $\gamma_i = 0.1$ (the lowest values we could study) and decreased to 1 $\mu$m for $\gamma_i \geq 1.2$. These data are consistent with a logarithmic decrease in $\Delta h^*$ as a function of $\gamma_i$, at least up to $\gamma_i = 1.2$. Beyond $\gamma_i = 1.2$ dilation remains constant with increasing shear strain (Figure 2.6 inset).
A subset of experiments was conducted with 0.3 or 0.5-cm thick layers (Figure 2.6). For the thinner layers, layer dilation was about 3 $\mu$m for the lowest $\gamma_l$ values and decreased to 2 $\mu$m for $\gamma_l = 1$. These data are consistent with the idea that shear is distributed across the entire layer thickness at low strains and then becomes localized, to a thickness that is independent of $h$, for larger strains. Shear localization and fabric development also influence the rheology of sheared granular layers [Rathbun et al., 2008]. In the next section, we extend the investigation of shear localization and consider results from tests conducted at constant macroscopic strain rate (e.g. Figure 2.3).

### 2.3.3 Slip velocity step tests

In addition to velocity step tests conducted after creep-mode loading (e.g. Figure 2.3) we ran dedicated experiments at controlled shear velocity, which included stepwise increases and decreases in velocity between 10 $\mu$m/s and 30 $\mu$m/s (Figure 2.7). These experiments were done with 10-mm thick layers and were designed to assess variations in rate/state friction parameters as a function of strain. Such variations have been used as a proxy for fabric development in sheared layers [Marone and Kilgore, 1993; Beeler et al., 1996]. The shear displacement at each velocity was 450 $\mu$m or 550 $\mu$m (Table 2.1). Velocity steps continued until a maximum displacement of $\sim$ 35 mm corresponding to shear strains of 3.5 to 4. During the initial phase of shear stress increase, velocity steps were partially obscured by non-linear strain hardening (Figure 2.7).
After friction reached steady state, we imposed step changes in load point velocity, which caused variations in sliding friction (inset to Figure 2.7). Upon an increase (decrease) in loading velocity, friction increased (decreased) and then decayed to a new steady over a critical slip distance, $D_c$ (Figure 2.7). The dependence of friction on slip rate and state (recent slip history) can be described by the rate and state friction relation:

$$
\mu = \mu(V, \theta) = \mu_0 + a \ln\left(\frac{V}{V_0}\right) + b \ln\left(\frac{V_0 \theta}{D_c}\right)
$$

and one of two evolution laws for the friction state variable [Dieterich, 1979; Ruina, 1983]:

$$
\frac{d\theta}{dt} = 1 - \frac{V_0 \theta}{D_c} \quad \text{(Dieterich Law)}
$$

$$
\frac{d\theta}{dt} = -\frac{V_0 \theta \ln\left(\frac{V_0 \theta}{D_c}\right)}{D_c} \quad \text{(Ruina Law)}
$$

where $\mu$ is the friction coefficient, $\mu_0$ is friction at a reference velocity $V_0$, $V$ is the sliding velocity, $\theta$ is a state variable, and $a$ and $b$ are dimensionless constants (Figure 2.7 inset). The friction parameters $a$, $b$, and $D_c$ are obtained by solving the coupled equations (2) and (3a) or (3b) along with a description of elastic interaction with the testing machine:

$$
\frac{d\mu}{dt} = k(V_l - V),
$$

where $k$ is apparatus stiffness divided by normal stress and $V_l$ is load point velocity [e.g. Marone, 1998].

In our experiments, we observed that a step increase in loading velocity causes a rapid increase in shear stress. The rate of stress increase with load point
displacement is equal to the system stiffness (Figure 2.7). At some point, the stress becomes sufficient to cause further slip within the layer and then frictional strength reaches a maximum, after which it decays to a new steady value (Figure 2.7 inset). The e-folding distance required to establish the new steady state sliding friction is the critical slip distance \(D_c\). We observe that decreases in loading rate cause a sudden drop in frictional stress, followed by strengthening to a new steady state level. When the steady-state change in friction has the same sign as the velocity change, such as shown in Figure 2.7, the material is said to exhibit velocity strengthening frictional behavior. Velocity weakening frictional behavior is defined by a lower steady state friction value at higher sliding velocity. Friction rate dependence is given by:

\[
a - b = \frac{\Delta \mu}{\Delta \ln V}.
\]

Previous works on granular and clay fault gouge have shown that negative values of the friction rate parameter, \(a-b\), are associated with localized shear [Marone \textit{et al.}, 1990; 1992; Beeler \textit{et al.}, 1996]. As fabric develops and shear becomes more localized the critical slip distance decreases [Marone and Kilgore, 1993]. According to Marone and Kilgore [1993] the reduction in \(D_c\) occurs because a smaller fraction of the bulk layer and fewer particle-particle contacts are contributing to shear.

\textbf{2.3.4 Evolution of the critical slip distance}

We analyzed velocity step tests to assess evolution of constitutive parameters with strain and fabric development (Figure 2.7). A non-linear, least-squares inversion method was used to obtain parameters, following the procedures of Blanpied \textit{et al.} [1998]. Each velocity step was modeled separately (Figure 2.8). In a few cases, the
data exhibit a small overall trend of strain hardening (or weakening), which we accounted for by including a linear term in the model. The best-fit model and a sensitivity analysis for the critical slip distance, $D_c$, are presented for two representative velocity steps in Figure 2.8 using both the Dieterich state evolution law (Equation 3a) and the Ruina evolution law (Equation 3b). For the velocity increase, the best-fit parameters are: $a = 0.0116$, $b = 0.0106$, and $D_c = 95 \mu m$ using the Dieterich law and $a = 0.0121$, $b = 0.0102$ and $D_c = 115 \mu m$ using the Ruina law (Figure 2.8). For the velocity decrease, the best-fit parameters are $a = 0.0137$, $b = 0.0117$, and $D_c = 152 \mu m$, and $a = 0.0131$, $b = 0.0110$ and $D_c = 108 \mu m$ for the Dieterich and Ruina laws, respectively (Figure 2.8). For reference, we fix the values of $a$ and $b$ in each case and compute three forward models using different values of $D_c$. Changing the value of $D_c$ by as little as 25 \mu m results in a significant misfit (Figure 2.8).

Comparison of forward models with similar values of $D_c$ shows that the friction behavior is asymmetric for velocity increases and decreases when analyzing the steps with Dieterich evolution. The value of $D_c$ is nearly a factor of 2 higher for velocity decreases compared to increases. Whereas the values of $D_c$ are symmetric when the data are fit using the Ruina law. There is significant covariance between parameters [e.g. Blanpied et al., 1998], but we focus here on $D_c$ – for fixed values of $a$ and $b$– to assess asymmetry and differences between velocity increases and decreases.

Using the procedure shown in Figure 2.8, we assess evolution of constitutive behavior as a function of shear strain by fitting velocity steps for multiple experiments. Values of $a-b$ are similar for velocity increases and decreases, with both
showing velocity strengthening and a slight reduction in magnitude for $\gamma < 2$ (Figure 2.9). The average value of $a-b$ for velocity increases is $0.0023\pm 0.0014$ compared to $0.0028\pm 0.0019$ for velocity decreases with the Dieterich Law. Using the Ruina Law these values change slightly to $0.0022\pm 0.0014$ and $0.0024\pm 0.0016$ for increases and decreases, respectively. This consistency is expected because $a-b$ represents a steady state response, which is independent of the details of state evolution. Our measurements are consistent with previous results for this material [Rathbun et al., 2008], which show velocity strengthening frictional behavior for normal stresses from 50 kPa to 5 MPa and slip velocities from 1 $\mu$m/s to 300 $\mu$m/s.

Considering the range of our data, which start at a shear strain of about 1, the critical slip distance is independent of shear strain, within the scatter in the data. However, $D_c$ is systematically different for velocity increases and decreases (Figure 2.9A, 2.9C) when using the Dieterich law. Mean values of $D_c$ are $93.3\pm 20.2$ $\mu$m and $182.9\pm 41.1$ $\mu$m for velocity increases and decreases, respectively in experiment p1572. These data and the sensitivity analysis of Figure 2.8 show a clear asymmetry in the critical slip distance for step increases and decreases in velocity. Friction approaches steady-state within a shorter slip distance after velocity increases than velocity decreases.

We modeled the same velocity steps with the Ruina state evolution law and find that the values of $D_c$ are symmetric for velocity increases/decreases (Figure 2.9). In experiment p1572, the mean for $D_c$ is $122\pm 53$ $\mu$m and $140\pm 18$ $\mu$m for velocity decreases and increases, respectively. In p1345 increases have a $D_c$ of $139.1\pm 34.7$ $\mu$m and decreases $131\pm 19$ $\mu$m when using the Ruina law. For experiment p1507 the mean
for increases is 123±29 μm and 121±13. There is no clear asymmetry within the scatter in these data.

In all cases the Dieterich law produces significant asymmetry for velocity increases/decreases. This is consistent with expectation, because the Dieterich law assumes that steady state is reached within a critical time; thus the slip that accumulates during that time should be larger for velocity increases than for velocity decreases. This would predict larger values of $D_c$ for velocity increases than decreases, which is opposite to what we observe (Figure 2.9). This issue is discussed further below.

2.3.5 Strain markers and localized deformation

Thin zones of blue sand were added to select experiments (Table 2.1) to track the strain distribution within the layer. These layers were carefully recovered after shear, impregnated with a low viscosity epoxy, and cut to expose a plane perpendicular to the layer and parallel to the shear direction (Figure 2.10). Photomicrographs in reflected light document offset of the markers and a combination of pervasive and localized strain (Figure 2.10). These images confirm that shear occurred within the sample and not at the interface with the rough forcing blocks. Strain markers are arcuate near the layer edges and curve into a boundary parallel (Y) orientation toward the center of the layer (Figure 2.10B). The thin zone of shear marker seen throughout the layer indicates that strain does not localize into a true Y shear, but into a narrow zone in the center of the sample. This implies that the
boundary parallel ‘Y shears’ in these experiments have finite width and that shear within them is not on an infinitesimally-thin plane.

Curvature of the markers along the layer boundaries indicates progressive localization with increased macroscopic strain (Figure 2.10B-D). Three transects were taken across the sheared sample, one at each boundary and one in the center of the shear marker (Figure 10C). The angular shear strain,

$$\gamma_a = \tan \psi,$$

where $\psi$ is the angle between the initial position of the shear marker and current position, was calculated in the curved portion of the marker using the method of [Ramsay and Graham, 1970]. The $\gamma_a$ can be calculated between each point along the transect (Figure 2.10D). We may compare angular shear strain, $\gamma_a$, to bulk shear strain across the layer, $\gamma$ (Equation 1). Bulk shear strain represents a macroscopic average whereas $\gamma_a$ can be used to infer strain in a localized area of the sample and may have values much larger than $\gamma$. We only present calculations of $\gamma_a$ along the curved portion of the strain marker, near the layer boundary (Figure 2.10C,D). In the central, high strain portion of the layer, Riedel shears and indentations of large grains into our shear marker preclude calculations of $\gamma_a$.

Figure 2.10D presents $\gamma_a$ as a function of position within the sample. Near the shear zone boundary $\gamma_a$ is near zero and increases toward the center of the layer. Strain is high in the central zone and local variations associated with large grains and slip surfaces make it difficult to resolve the peak strain value. Thus, Equation (6) does not give an accurate assessment of strain in the central region. Nevertheless, the overall strain distribution can be approximated with a normal distribution and
compared to measurement of macroscopic layer strain, $\gamma$, from the experiment. We integrate the normal distribution to derive a total shear displacement of 16.6 mm. The value can be compared to the measured shear displacement imposed at the layer boundaries, which was 30.4 mm, and the bulk shear strain, which was 3.9. We may assume that the slip derived from local angular shear strain, 16.6 mm, represents only that which occurred outside the central zone of high strain (Figure 2.10). This amount of slip corresponds to the outer, curved portion of the shear marker. In this case, the remaining $\gamma$ of 1.9 would occur in the central, boundary parallel section, which is roughly 1.6 mm thick and in the center of the sample. The 1.6 mm thickness corresponds to a few grain diameters in thickness. To account for the remaining $\gamma$ in the layer, a peak shear strain of 8.6 is required for the highly localized section near the center of the layer, which is reasonable.

Shear markers and the spatial distribution of strain in the layer show that shear is initially pervasive and becomes localized (Figure 2.11). One possibility is that Y shear formation could simply cut the markers as shown in Figure 2.11C. However, the photomicrographs document significant curvature of the markers as expected for spatially-progressive localization (e.g. Figure 2.11D).

We posit that boundary parallel shear localization begins on multiple surfaces and progresses to a certain point before one or more of the zones coalesce to form a master shear band. Our observations suggest that during the initial stages of localization; shear surfaces nearest the layer boundary arrest first, while those nearer to the center continue to shear. The low relative amounts of strain on the boundary,
progressing to larger relative amounts of strain near the center of our sample causes a curvature of the strain markers (Figure 2.10).

2.4. DISCUSSION

The results of this study document strain localization and systematic changes in frictional behavior as a function of accumulated shear strain. Creep-mode tests and perturbations in shear stress level show consistent layer dilation for an increase in shear stress, and we use dilation as a proxy for shear localization within the layer. These results are consistent with previous works, but we add to those and extend the investigation to show how progressive fabric development affects frictional behavior. Our work on slip velocity perturbations compares velocity increases and decreases, and investigates symmetry in the frictional behavior.

2.4.1 Dilation as a proxy for shear localization

For granular materials, dilation occurs when shear-induced grain rearrangement causes a local increase in porosity [Reynolds, 1885; Mead, 1925]. Our measurements of macroscopic layer dilation form the basis for assessing shear localization and the relationship between fabric development and frictional behavior. We perform two tests of the hypothesis that layer dilation is a valid proxy for shear localization. These involve: 1) experimentally varying initial layer thickness (Figure 2.6), and 2) using strain markers to document slip distribution within the layers (Figure 2.10). In addition, we can compare our dilation data to inferences about localization based on friction constitutive parameters (Figure 2.9).
We varied initial layer thickness from 3 mm to 1 cm. Data for 1 cm thick layers show decreasing dilation with increasing shear strain (Figure 2.6). If this decrease is the result of shear localization, layer thickness will influence dilation at low strain, when deformation is pervasive, but at high strain, once deformation is localized, layer thickness will have no influence on dilation. Experiments on 0.3 and 0.5 cm layers produce half the dilation of 1 cm thick layers at $\gamma \sim 0.15$ (Figure 2.6), which is consistent with dilation throughout the layer (e.g. distributed shear). At $\gamma \sim 0.25$ our data are less convincing (Figure 2.6 inset). The 0.5-cm thick layers fall below the line defined by the 1 cm layers, but the data for 1-cm thick layers have large uncertainty (Figure 2.6 inset). At higher strains, when $\gamma = 1$, all layer thicknesses show equal dilation within experimental uncertainty and reproducibility. The large variability of dilation at $\gamma$ suggests that localization may be complete at a slightly larger value than 1. These data are consistent with the hypothesis that deformation has localized into the same effective thickness for all macroscopic layer thicknesses. The micrographs and strain markers also support the conclusion that shear strain becomes localized within the layer for macroscopic shear strains in the range $\sim 1-2$.

Friction constitutive parameters have been used as a proxy for shear localization [e.g. Marone and Kilgore, 1993; Beeler et al., 1996; Scruggs and Tullis, 1998; Mair and Marone, 1999; Frye and Marone, 2002; Mitchell and Soga, 2005]. Marone and Kilgore [1993] sheared layers of granular and powdered quartz and found that the critical slip distance $D_c$ decreased until $\gamma \sim 6$. Mair and Marone [1999] investigated a range of normal stresses and slip rates and found that $a-b$ evolves until
$\gamma \sim 4$. Beeler et al. [1996] conducted ring shear experiments on granular granite and found continued evolution of $a$-$b$ up to $\gamma > 50$.

In our experiments evolution of the friction constitutive parameters appears to be complete by $\gamma \leq 1$-$2$. The maximum $\gamma$ we impose is $\sim 4$, which is lower than other studies. Also, our material is a glacial till, with inherent heterogeneity of grains and a large size distribution. We believe this is part of the cause of the scatter in $D_c$ as well as $\Delta h^*$ measurements. The values of $a$-$b$ display a clear evolution and decreasing trend until $\gamma \sim 2$ (Figure 2.9), consistent with localization assessed from the strain markers (Figure 2.10C). It is possible that continued strain would lead to further reduction of $a$-$b$. In the studies of Marone and Kilgore [1993], Beeler et al. [1996] and Mair and Marone [1999] most of the evolution of $a$-$b$ takes place at low strain, consistent with both our friction constitutive and creep dilation data.

Laboratory studies of till localization using anisotropy of magnetic susceptibility (AMS) fabric indicate that that localization evolves until $\gamma$ on the order of 100 [Larsen et al., 2006; Thomason and Iverson, 2006; Iverson et al., 2008]. Unfortunately these studies do not include data on the friction constitutive parameters for sheared till, and are all conducted in ring shear apparatuses. This configuration has a much lower stiffness than our apparatus and is typically used at lower normal stress. Thus, these tests require higher shear strain to reach steady-state sliding friction, which makes direct comparison difficult. Past experiments on the tills used by Thomason and Iverson [2006] and Iverson et al. [2008] indicate velocity-weakening behavior, but without information on $D_c$ or the evolution of friction constitutive parameters with strain [e.g. Iverson et al., 1998]. In general, our results
are consistent with these studies. We see that strain markers initially rotate in a manner consistent with distributed deformation and then record progressive localization in a narrow band of finite width, before slip localizes onto a surface that offsets the markers (Figure 2.10). The edges of the strain markers that were rotated during distributed deformation show both curvature and thinning toward the center of the shear zone, indicating that the transition to localized shear occurred progressively. It is possible that grains in our experiments continue to rotate and align into a higher-order preferred orientation than traditional shear fabrics. However, clast rotation and particle alignment are beyond the scope of this study and we do not attempt to verify our results using preferred axis orientation.

Our results are consistent with those of Logan et al. [1992], who sheared granular and clay-rich layers in the triaxial, sawcut configuration. They report pervasive deformation and the formation of oblique, Riedel shears during the initial, hardening portion of the stress-strain curve, followed by formation of boundary parallel Y-shears as frictional strength approaches steady-state. Scruggs and Tullis [1998] also document localized shear in a boundary parallel zone within layers of mica and feldspar. They observe velocity weakening frictional behavior and make a connection with shear localization and possible stick-slip instability.

Strain markers in our experiments indicate that Y-shears have finite thickness and that they begin to form before the peak frictional strength (Figure 2.12). We find that dilation begins to decrease at $\gamma \sim 0.15$, which is before the peak shear strength (Figure 2.12). Dilation continues to decrease as the shear stress transitions to a steady sliding strength at $\gamma \sim 0.3$ in most experiments (Figure 2.12B). By $\gamma \sim 1.2$, the
decrease in dilation is complete and sliding friction is steady. During the decrease in dilation, the mode of deformation changes from a distributed model to one in which most shear occurs in a boundary parallel zone in the center of the sample.

2.4.2 Symmetry of frictional behavior for velocity increases and decreases

In the context of rate and state friction, the two main state evolution laws (Equations 3a and 3b) make different predictions regarding the symmetry of the response to velocity increases and decreases. The Ruina law predicts symmetric behavior while the Dieterich law predicts larger values of $D_c$ for velocity increase than for velocity decreases. We model our results with both laws and find complex behavior. Our experiments show a clear asymmetry in the frictional response to velocity perturbations when data are fit using the Dieterich state evolution law. Moreover, the asymmetry is opposite to that predicted for the Dieterich law. The critical slip distance for velocity decreases are a factor of 2 larger than those for velocity increases (Figure 2.9), whereas they should be smaller, according to standard interpretation [e.g., Marone, 1998].

Many previous studies of the evolution of rate/state friction with strain have focused on only velocity increases or decreases, without considering the question of symmetry. The experiments by Marone and Kilgore [1993] show decreasing $D_c$ with increased shear strain in layers of granular quartz. They analyzed velocity decreases in detail and noted a similar trend for velocity increases; however they did not compare velocity increases and decreases. The experiments of Mair and Marone [1999] find that $D_c$ increases with velocity as predicted by the Dieterich law. Marone
and Cox [1994] show that $D_c$ increases with displacement for roughened gabbro blocks due to the production of a gouge zone. Within the reproducibility of their data, $D_c$ is symmetric for velocity increases and decreases. Asymmetry of velocity increases and decreases has been observed in dilatancy produced by velocity steps [e.g. Beeler et al., 1996; Hong and Marone, 2005].

When data for velocity increases and decreases are compared directly, the frictional evolution we observe is indistinguishable for velocity increases and decreases (Figure 2.13A). This symmetry between the increase and decrease in velocity suggests that the Ruina law may be more appropriate for these data. This is in contrast to the work of Beeler et al. [1994] who preformed experiments on granite and quartzite and showed that frictional state evolved primarily as a function of time, as predicted by the Dieterich state evolution law. For comparison we present data from a second experiment showing asymmetry of the friction evolution (Figure 2.13B). For these data, pure quartz was sheared at a normal stress of 25 MPa. Velocity decreases appear to reach peak friction at smaller displacements then velocity increases and friction evolves over a longer distance.

Our data lend clear support for the Ruina law interpretation of frictional state evolution. However, an alternative hypothesis should be considered, given that we find evidence for shear localization in a gouge layer. One could argue that the Dieterich law is correct, and that changes in the degree of shear localization explain our data. Previous studies have established that the critical slip distance scales with particle size [e.g. Dieterich, 1981; Marone and Kilgore, 1993] in a manner consistent with Rabinowicz’s [1958] original interpretation of $D_c$ in terms of the lifetime of
adhesive contacts. Marone and Kilgore [1993] extended this idea and proposed that the critical slip distance for granular shear scales with the number of particles within a shear band (see also Marone et al. 2009). That is, for a zone of thickness $T$, the effective critical slip distance $D_{cb}$ is given by the sum of contributions from individual contacts within the zone:

$$D_{cb} = n D_c \chi$$  \hspace{1cm} (7)

where $\chi$ is a geometric factor to account for contact orientation and $n$ is the number of particle contacts in the shear band.

For a fault zone of thickness $T$, the effective critical slip distance represents contributions from each contact within the zone. Particle diameter $d$ can be related to $D_c$ via contact properties as:

$$D_c = d \zeta$$  \hspace{1cm} (8)

where $\zeta$ is a constant including elastic and geometric properties and the slip needed for fully-developed sliding at the contact [Boitnott et al., 1992]. Combining these relations and the constants, yields a relation between $D_{cb}$ and shear band thickness:

$$D_{cb} = T \gamma_c$$  \hspace{1cm} (9)

where $\gamma_c$ is the critical strain derived from slip increments on individual surfaces within the shear zone [Marone and Kilgore, 1993; Marone et al., 2009].

In the context of this model, asymmetry in frictional behavior for velocity increases vs. decreases can be explained by dynamic variation in shear band thickness. Larger values of the effective critical slip distance imply that a larger number of contacts, and a thicker shear band, are actively slipping. Thus, we posit that a transient increase in the imposed shearing rate causes contacts to strengthen, via
the friction direct effect, followed by weakening. This would cause a transient widening of the shear band, as interparticle slip was temporarily arrested and particles rotated, with attendant local dilation. With continued slip, and contact weakening via the friction evolution effect, the shear band would thin. We assume that interparticle contacts undergo velocity weakening, even where the macroscopic friction response is velocity strengthening [e.g. Marone et al., 1990], and thus the shear band thins as slip is focused on weaker contacts. On the other hand, a transient decrease in the imposed shearing rate has the effect of, effectively, delocalizing shear by equalizing age (frictional state) of interparticle contacts within the active shearing zone relative to those outside the zone. This growth (thickening) of the actively shearing zone as it incorporated more material from ‘spectator’ regions [e.g. Mair et al., 2002; Mair and Abe, 2008] would produce larger $D_{cb}$.

2.4.3 Localization and till rheology

The progressive localization that we observe is consistent with the changes in till rheology noted by Rathbun et al. [2008]. They observe a gradual change from a pseudo-viscous rheology with a stress exponent of order 1 at $\gamma \sim 0$, to values $> 30$ as a function of accumulated shear strain. Our experiments were conducted using the same experimental configuration and on the same material as those of Rathbun et al. [2008]. The change in the stress exponent and transition of the stress exponent occurs over the same interval as our decrease in $\Delta h^*$. Our data support the idea that the rheology of till is a function of shear localization. These results suggest that as long as shear remains localized in till a frictional (Coulomb-plastic) rheology is appropriate.
rather than a viscous rheology. In the case that pore pressure or some other feedback destroys localization, a viscous rheology could apply until shear strain accumulates.

2.5 CONCLUSIONS

Laboratory experiments on a saturated, mixed grain-size granular media show that shear begins to localize prior to the peak frictional strength. Dilation of the layer in creep experiments and the evolution of the friction constitutive parameters (a-b and critical slip distance) all show a progressive transition from distributed shear to localized deformation. Passive strain markers in the layer confirm that distributed deformation occurs at the lowest shear strain, and then deformation occurs in a localized zone in the center of the sample. This localization occurs progressively starting at a strain of ~0.15 and continues after the peak strength until a shear strain of ~1-2. The localized zone is similar to boundary parallel Y-shears in the sample. Mapping shear strain across of the deforming layer using embedded shear markers shows that the formation of the boundary parallel layer occurs at shear strain < 1.9. These localization features occur despite the material showing velocity strengthening frictional behavior at all shear strains. Step increases and decreases in velocity indicate an asymmetry in some of the rate and state friction constitutive parameters. The asymmetry is observed in the critical slip distance, while the velocity dependence of friction remains constant for velocity increases and decreases when the data are modeled with the Dieterich law. If the data are modeled using the Ruina law, asymmetry is not observed or is obscured by scatter in the data owing to the heterogeneous nature of the material. Since the velocity dependence of friction
remains constant for increases and decreases in velocity, asymmetry could be the result of changes in the degree of shear localization within the bulk layer. However, direct comparison of velocity increases and decreases shows nearly identical evolution of the critical slip distance, leading us to favor the Ruina law for friction state evolution.

The results of this study imply that localization of strain can occur at low shear strains in preexisting faults and glacial deformation zones. The bulk rheology of subglacial till may be the result, and not the cause, of localization in the shearing till layer. Highly localized zones within fault gouge such as the principle slip surface may begin to form at low shear strain, regardless of velocity weakening or strengthening behavior.

2.6. ACKNOWLEDGMENTS

This work was funded by: ANT-0538195. We thank Hitoshi Banno and Mike Underwood for providing XRD analysis. Particle size analysis was conducted at the Materials Research Institute, at Penn State. This manuscript was greatly improved by the comments of two anonymous reviews, Associate Editor Dan Faulkner and comments on early versions by André Niemeijer and Sébastien Boutareaud.

REFERENCES


Mead, W. J. (1925), The geologic role of dilatancy, *J. Geol.* 33, 685-698.


Scott, D., C. Marone, and C. Sammis (1994), The apparent friction of granular fault
Scruggs, V.J. and T.E Tullis (1998), Correlation between velocity dependence of
friction and strain localization in large displacement experiments on feldspar,
1169-1178.
Storti, F., A. Billi, and F. Salvini (2003), Particle size distributions in natural
Let.* 206, 173-186.
Thomason, J. F., and N. R. Iverson (2006), Microfabric and microshear evolution in
Truffer, M., W. D. Harrison, and K. A. Echelmeyer (2000), Glacier motion
dominated by processes deep in underlying till, *J. Glaciol.* 46(153), 213-221.
Tsai, V. C., J. R. Rice, and M. Fahnestock (2008), Possible mechanisms for glacial
Underwood, M.B., N. Basu, J. Steurer and S. Udas (2003), Data report: 
Normalization factors for semi-quantitative X-ray diffraction analysis, with
application to DSDP Site 297, Shikoku Basin, in *Proceeding of the Ocean
Drilling Program, Scientific Results* 190/196, edited by Mikada, H., et al.,
Ocean Drilling Program: College Station, Texas, 1-26.
Winberry, J. P., S. Anandakrishnan, R. B. Alley, R. A. Bindschadler, and M. A. King
(2009), Basal mechanics of ice streams: Insights from the stick-slip motion of
Whillans Ice Stream, West Antarctica, *J. Geophys. Res.*, 114, F01016,
<table>
<thead>
<tr>
<th>Experiment</th>
<th>$\sigma_0$ (MPa)</th>
<th>Exp. Type</th>
<th>$h_i$ (mm)</th>
<th>$\gamma_i$</th>
</tr>
</thead>
<tbody>
<tr>
<td>p729</td>
<td>1</td>
<td>Creep- 2% steps</td>
<td>10</td>
<td>1.12</td>
</tr>
<tr>
<td>p731</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>1.15</td>
</tr>
<tr>
<td>p732</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>1.12</td>
</tr>
<tr>
<td>p737</td>
<td>1</td>
<td>Creep- 2% steps</td>
<td>10</td>
<td>1.13</td>
</tr>
<tr>
<td>p746</td>
<td>1</td>
<td>Creep- 2% steps</td>
<td>10</td>
<td>1.19</td>
</tr>
<tr>
<td>p750</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>1.14</td>
</tr>
<tr>
<td>p757</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>0.1</td>
</tr>
<tr>
<td>p760</td>
<td>1</td>
<td>Creep- 2% steps</td>
<td>10</td>
<td>0.18</td>
</tr>
<tr>
<td>p761</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>0.16</td>
</tr>
<tr>
<td>p1125</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>0.15</td>
</tr>
<tr>
<td>p1131</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>0.25</td>
</tr>
<tr>
<td>p1167</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>0; 2.65</td>
</tr>
<tr>
<td>p1194</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>0.19; 1.78</td>
</tr>
<tr>
<td>p1215</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>5</td>
<td>0.25</td>
</tr>
<tr>
<td>p1216</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>5</td>
<td>0.95</td>
</tr>
<tr>
<td>p1228</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>0.56</td>
</tr>
<tr>
<td>p1229</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>5</td>
<td>0.3</td>
</tr>
<tr>
<td>p1230</td>
<td>5</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>0.87</td>
</tr>
<tr>
<td>p1231</td>
<td>5</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>0.3</td>
</tr>
<tr>
<td>p1253</td>
<td>0.5</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>1.05</td>
</tr>
<tr>
<td>p1263</td>
<td>0.5</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>0.09</td>
</tr>
<tr>
<td>p1345</td>
<td>1</td>
<td>V-Steps- 10-30 μm/s</td>
<td>10</td>
<td>N/A</td>
</tr>
<tr>
<td>p1494*</td>
<td>1</td>
<td>Const. Disp.</td>
<td>10</td>
<td>N/A</td>
</tr>
<tr>
<td>p1507</td>
<td>1</td>
<td>V-Steps- 10-30 μm/s</td>
<td>10</td>
<td>N/A</td>
</tr>
<tr>
<td>p1508*</td>
<td>1</td>
<td>V-Steps- 10-30 μm/s</td>
<td>10</td>
<td>N/A</td>
</tr>
<tr>
<td>p1513</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>0.52</td>
</tr>
<tr>
<td>p1572</td>
<td>1</td>
<td>V-Steps- 10-30 μm/s</td>
<td>10</td>
<td>N/A</td>
</tr>
<tr>
<td>p1814</td>
<td>1</td>
<td>Creep- 7.5% steps</td>
<td>10</td>
<td>0.24; 1.27</td>
</tr>
<tr>
<td>p1824</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>1.07</td>
</tr>
<tr>
<td>p1906</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>1.17</td>
</tr>
<tr>
<td>p1910</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>3</td>
<td>0.7</td>
</tr>
<tr>
<td>p1938</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>0.17; 0.56</td>
</tr>
<tr>
<td>p1940</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>3</td>
<td>0.15; 0.92</td>
</tr>
<tr>
<td>p1942</td>
<td>1</td>
<td>Creep- 5% steps</td>
<td>10</td>
<td>4.3</td>
</tr>
</tbody>
</table>

Table 2.1. Experiment details. *denotes experiments that included strain markers. $h_i$ is initial layer thickness. $\gamma_i$ is initial shear strain.
Figure 2.1. Double-direct shear configuration. The nominal frictional contact area was 10 cm x 10 cm and did not change with shear. Initial layer thickness was 0.3, 0.5 or 1 cm. The block arrangement was surrounded by a rubber membrane and filled with water. The reservoir was open to the atmosphere leading to saturated, drained conditions.
Figure 2.2. Grain-size distribution for Caesar till. Samples were air-dried and sieved. Fine fraction (< 63 μm) was analyzed using Laser Diffraction. Experiments are conducted on all grains that passed through a 1 mm sieve.
Figure 2.3. Complete stress-strain history for a representative experiment. Layers were sheared under controlled shear stress or constant shear displacement rate (boxed areas). (A) Shear stress and strain vs. time to highlight creep mode section. (B) Shear stress vs. shear strain for the same experiment. The preconditioning shear strain, $\gamma_i$ was varied systematically to study the effect of shear localization on creep behavior. We measured $\tau_{res}$ during the (preconditioning) run-in, prior to creep tests under stress control. After completion of creep testing, we switched back to constant shear displacement rate. Modified from Rathbun et al. [2008].
Figure 2.4. Layer thickness data for two representative shear stress steps during creep loading. Dashed line shows raw layer thickness. Lower line shows change in layer thickness, $\Delta h^*$ corrected for geometric thinning. Note rapid dilation and steady-state change in thickness caused by stress steps. (Inset) temporal evolution of corrected layer thickness during one complete stress step.
Figure 2.5. Scaling of dilation $\Delta h^*$ with magnitude of shear stress change. Note that $\Delta h^*$ increases with stress step size for low shear strain ($\gamma_l < 0.2$) but not for higher shear strain. High strain corresponds to $\gamma_l = 1-1.2$ for 0.01 MPa and 0.025 MPa steps and $\gamma_l = 1.28$ for 0.045 MPa steps. Low strain corresponds to $\gamma_l < 0.02$ for 0.01 MPa and 0.025 MPa steps and $\gamma_l = 0.026$ for 0.045 MPa steps.
Figure 2.6. Layer dilation $\Delta h^*$ induced by shear stress perturbations of magnitude 0.05 $\tau_{res}$ (e.g. Figure 2.4) plotted versus $\gamma_i$, the initial shear strain prior to creep tests (e.g. Figure 2.3). Inset shows data versus the log of $\gamma_i$. Dilation is largest at low $\gamma_i$, where shear is expected to be pervasive, and decreases systematically as a function of $\gamma_i$. The 0.5-cm thick layers show about half the dilation of 1-cm thick layers at low strain, ~3 $\mu$m and 6 $\mu$m, respectively, but the values are about the same for a shear strain of 1. Each data point represents several stress steps. Error bars show one standard deviation from the mean and data with error $> 30\%$ of the mean are not plotted.
Figure 2.7. Controlled slip velocity experiment. Velocity is stepped between 10-30 μm/s and held for either 450 μm or 550 μm for the entire displacement range of the apparatus. Each velocity step causes a spike, then decay in friction. (Inset) Frictional response to a step increase in velocity. When velocity is increased, frictional strength increases instantaneously and then decays over a characteristic slip distance ($D_c$). In this example, the instantaneous increase is larger than the subsequent decay, and therefore steady state sliding friction exhibits positive $a-b$ and velocity-strengthening behavior.
Figure 2.8. Two velocity steps with velocity increased (A, B) and decreased (C, D) for experiment p1572. (A) Velocity is instantaneously increased from 10 μm/s to 30 μm/s. After the increase, friction increases then decays to a new steady value over the critical slip distance ($D_c$). Inverse modeling using the Dieterich law (thick, grey line) produces values of 0.0116, 0.0106, and 94.7 μm for the parameters $a$, $b$, and $D_c$, respectively. Forward models with different values for the critical slip distance are shown for reference. (B) Analysis using the Ruina law. Inverse modeling produces values of 0.0121, 0.0102, and 114.7 μm for $a$, $b$, and $D_c$, respectively. (C) Velocity is decreased to 10 μm/s and modeled with the Dieterich law. Values are 0.0137, 0.0117, and 151.5 μm for $a$, $b$, and $D_c$, respectively. Reference forward models are shown for different critical slip. (D) Velocity decrease modeled with the Ruina law. Values are 0.0131, 0.0110, and 108.0 μm for $a$, $b$, and $D_c$, respectively.
Figure 2.9. Friction constitutive parameters for two select experiments. Velocity decreases and increases are presented for the Dieterich law (A, C) and Ruina law (B, D). The velocity dependence of friction \((a-b)\) decreases with increased shear strain in all cases. Values for \(a-b\) show no variation between velocity increases or decreases when using either law. Values of the critical slip distance are dependent on the sign of the velocity step when modeled with the Dieterich Law (A, C). Modeling the data with the Ruina law produces equivalent \(D_c\) for both velocity increases and decreases.
Figure 2.10. Photomicrograph showing shear strain within the layer. Total bulk shear strain in the experiment was 3.9. (A) The strain markers were initially vertical; arrows on the top and bottom indicate sense of shear. (B) The 3 sets of strain markers are outlined. Due to the large displacement in the sample, only the central marker is complete. During disassembly, the sample parted on the left most strain marker. The shear marker curves into a boundary parallel zone in the center of the sample. Fiducial lines represent the angular shear strain from the rotation of a vertical line. (C) Enlarged image of (A) with three transects across the shear zone. The angular shear strain, $\gamma_\alpha$, is calculated between each point and presented in (D). (D) Angular shear strain as a function of position across the sample. Data symbols and position correspond to (C) and the thickness of the sample. The displacement from the overlaid normal distribution (solid black line) corresponds to bulk shear strain, $\gamma$, $\sim$ 2 leaving $\gamma \sim$ 1.9 in the center 1.6 mm of the sample. Micrograph from experiment p1508; no vertical exaggeration.
Figure 2.11. Idealized strain distribution for localized and distributed deformation. (A) Layer before shear with two theoretical strain markers. (B) During distributed deformation passive strain markers are continuously offset in simple shear. (C) After initially distributed deformation, markers are offset by localized deformation in a thin zone or plane. (D) Distributed deformation transitioning into localized strain. Boundary zones stop contributing to layer deformation and neighboring grains continue to shear resulting in a slight offset of the marker. This process continues causing an apparent curvature of the strain marker until all shear occurs in the center of the sample.
Figure 2.12. Layer dilation data with corresponding stress-strain curves for experiments at a normal stress of 1 MPa and initial layer thickness of 1 cm. Dilation begins at $\gamma_i \sim 0.15$ and continues until $\gamma_i \sim 1$. During the low strain portion distributed deformation is occurring throughout the layer. During the decrease in dilation shear is progressing into the localized state. This transition begins before the peak frictional strength and continues until after steady sliding friction has been achieved. Grey scale represents the progression from distributed shear (grey) to localized deformation (white).
**Figure 2.13.** (A) Velocity increases and decreases for experiment p1345 on Caesar till at a normal stress of 1 MPa. Seven steps are shown: three velocity increases (black) and four decreases (red). Velocity decreases are inverted (green) for comparison. Peak friction, and the evolution to steady-state are indistinguishable for increases and inverted decreases. (B) Velocity increases and decreases for pure quartz samples shear at a normal stress of 25 MPa. Velocity increases and decreases exhibit differing behavior with the decreases reaching peak friction at smaller displacements and evolving to steady state and larger slip. Peak friction is slightly small for decreases than increases.
Chapter 3. SYMMETRY IN RATE AND STATE

FRICTION

Andrew P. Rathbun$^{1,2}$

Chris Marone$^{1,2}$

$^1$Department of Geosciences, Penn State University, University Park, PA 16802

$^2$Penn State Sediment and Rock Mechanics Laboratory

To be submitted JGR
ABSTRACT

We performed experiments to investigate the symmetry/asymmetry of rate and state friction (RSF). Experiments were conducted in double-direct shear at 1 or 25 MPa normal stress with shear velocity stepped by a factor of 3 or 30. Experiments were conducted on one of three granular materials or on bare surfaces of Westerly granite; each of these materials has different frictional behavior. We find that the Ruina slip law, which predicts frictional symmetry between velocity increases and decreases, better matches our data than the Dieterich ageing law, which predicts that velocity decreases should evolve to steady state at smaller displacement. Conversely, some increases reach steady state friction at smaller displacement than decreases, an unexpected result from RSF. On bare surfaces the frictional response is symmetric. For factor of 30 velocity steps in granular materials, two distinct length scales are required to reach steady state. We hypothesize that asymmetry and two-state behavior is caused by changes in the localized shearing layer during the velocity step. In all cases dilation after a velocity increase is smaller than compaction after a decrease. In select experiments shear was stopped and normal stress was oscillated before repeating the velocity steps. In all cases we find that these oscillations decrease the critical slip distance, $D_c$. Reduction of $D_c$ reduces the stability of a shearing media, enhancing the possibility of seismic slip. Our experiments show that changes in the localized shearing zone of fault gouge can influence the RSF parameters that control earthquake rupture.
3.1. INTRODUCTION

Earthquakes in the brittle crust are controlled by frictional processes [e.g. Scholz, 2002] with observations from both exhumed faults [e.g. Logan et al., 1979; Chester and Chester, 1998; Cashman and Cashman, 2000; Faulkner et al., 2003; Hayman et al., 2004; Cashman et al., 2007] and faults at depth [Zoback, et al., 2010] indicating that earthquakes are hosted in gouge zones. Fault gouge is typically unconsolidated near the surface with the amount of lithification increasing with depth. The gouge zone can range in thickness from the micron to 10’s of meters scale [Scholz, 2002; Sibson, 2003].

Typically, brittle faulting is described using the rate and state friction (RSF) laws [Dieterich, 1979; 1981; Ruina, 1983]. In these empirical formalizations, second order friction variations are dependent on both slip velocity and a variable of state, thought to be the average contact lifetime between asperities [e.g. Rabinowitz, 1958], and the slip history of the sample. In its simplest form the RSF equation is

$$\mu(V, \theta) = \mu_0 + a \ln\left(\frac{V}{V_0}\right) + b_i \ln\left(\frac{V_0 \theta_i}{D_{ci}}\right) \ (i=1,2)$$  \hspace{1cm} (1)

where $\mu$ is sliding friction at a velocity, $V$, and $\mu_0$ is sliding friction at a reference velocity, $V_0$. The constitutive constant, $a$, often referred to the as the direct effect, is thought to be the result of Arrhenius processes resulting from breaking bonds at the atomic level [Rice et al., 2001]. The evolution effect, $b$, is related to the reference sliding velocity, a variable of state, $\theta$, and a critical length scale, $D_c$. Equation (1) can be written as either 1-state behavior when $i=1$ or two-state with $i=2$. In the case the two-state behavior, two evolution effects with two distinct length scales are present. In most cases only one length scale is considered and $b_2$ is taken to be 0. Two-state
behavior is thought to arise from complex shear fabric in the sample; however, two-
state variable behavior is usually ignored due to complications of interpreting two
length scales.

The rate and state friction equation is usually coupled with one of two
common evolution laws for the state variable:

\[
\frac{d\theta_i}{dt} = 1 - \frac{V\theta_i}{D_{ci}} \quad (i=1,2) \quad \text{Dieterich “aging” law} \quad (2)
\]

\[
\frac{d\theta_i}{dt} = -\frac{V\theta_i}{D_{ci}} \ln\left(\frac{V\theta_i}{D_{ci}}\right) \quad (i=1,2) \quad \text{Ruina “slip” law} \quad (3)
\]

These laws differ in their behavior at \( V=0 \). In this case, the aging law predicts
evolution of the state variable due to aging at grain-to-grain contacts, while the slip
law is undefined. In the aging law, contact lifetime is the dominant factor, while in
the slip law velocity and slip have a greater effect on state evolution.

The finite stiffness of the earth’s crust and experimental apparatuses requires
that the RSF equation (1) and the evolution laws (2, 3) be coupled with a term
describing the elastic stiffness of the loading apparatus. Written in terms of friction
[e.g. Gu et al., 1984] the change in stiffness with time is:

\[
\frac{d\mu}{dt} = k(V_l - V) \quad \text{(4)}
\]

where \( k \) is the elastic loading stiffness of the loading apparatus, \( V_l \) is the loadpoint
velocity and \( V \) is velocity at the slip surface.

Figure 3.1 presents two synthetic velocity steps with different \( k \), but the same
RSF parameters and velocities, Figure 3.1a is the Dieterich aging law and Figure 3.1b
is the Ruina slip law. Velocity increases with stiffness ranging \( k = 0.004 \mu m^{-1} \) to
are presented along with decreases. Both increases and decreases also include the theoretical case of $k = \infty$. The decreases are inverted for comparison to the increases and presented as a mirror image. At low stiffness, a pronounced difference exists between the peak friction for the velocity increases and decreases with the increasing having a greater amplitude (Figure 3.1). Also, there is a difference in a pre-peak behavior of velocity increases and decreases; the velocity decreases require a smaller slip to reach peak friction than increases, indicating a higher apparent stiffness. The differences in pre-peak behavior become less pronounced with increased stiffness, with no difference in the infinitely stiff material. Contrasting Figures 3.1a and 3.1b shows that the effects of stiffness are less pronounced for both peak friction and the displacement required for peak friction in the Ruina formulation. Also evident from Figure 3.1 is the different displacements required for the friction evolution of the velocity increases and decreases to match. For the Dieterich law (Figure 3.1a) the friction curves match starting at displacement $\sim 110 \, \mu m$ whereas for the Ruina law (Figure 3.1b) the increase/decrease curves begin to match at displacement $\sim 50 \, \mu m$. The Dieterich law’s prediction of time as the controlling variable in state evolution leads to steady state friction at a smaller relative displacement for velocity decreases versus increases, whereas in the Ruina formulation, slip is the controlling factor causing both decreases and increases to reach steady state at equal displacements (Figure 3.1).
3.1.1 Comparison of Evolution Laws

While similar in form, the different approach of each evolution law yields important distinctions in predictions of seismic behavior and at large velocity perturbations. For small velocity steps, the two evolution laws show unique, but not vastly different behavior (i.e. Figure 3.1). As the size of velocity steps increases the behavior of the laws diverges [i.e. Ampuero and Rubin, 2008]. Several investigators have noted that to reproduce Gutenberg-Richter phenomena and slip-pulses, contact aging, and thus the Dieterich law is required [e.g. Heaton, 1990; Rice, 1993; Perrin et al., 1995; Beeler and Tullis, 1996]. Ampuero and Rubin [2008] found that the nucleation length varies considerably depending on which law is used. They show that the nucleation zone for the Dieterich law should approach ~1 km, which could be observable while the nucleation zone when using the Ruina law is ~100x smaller.

Previous laboratory experiments have been inconclusive in separating which law should be used. Early experiments tended to favor the Ruina law due to its prediction of symmetry between velocity increases and decreases [e.g. Tullis and Weeks, 1986; Marone et al., 1990]. Imaging experiments by Dieterich and Kilgore [1994] showed that real contact area evolved with normal stress and time, favoring the Dieterich law. Beeler et al. [1994] preformed experiments in which the stiffness of the loading apparatus was varied. They found that the Dieterich law better fit their data over a range of hold times from ~3s to $10^5$ s. Blanpied et al. [1998] observed a better match with the Ruina law and two-state behavior in both room temperature and experiments up to 800 °C. In low-stress experiments (normal stress equal to 1 MPa) on glacial till Rathbun and Marone [2010] observed that modeling for the RSF
parameters yielded a longer $D_c$ for velocity decreases than increases, opposite Dieterich’s prediction; hence they favored the symmetric Ruina law. They also proposed an alternative model in which the shear zone changes in width for velocity increases and decreases. In such a model, interactions in the out-of-shear material or porosity change in the shear zone works to override the commonly used laws.

*Rathbun and Marone* [2010] proposed that changing shear zone width within a localized sample might cause samples that exhibit time-dependent effects to better match the slip law. They hypothesized that the shear zone widens or out-of-shear materials interact with the localized zone to lengthen the critical slip distance after a velocity decrease. In this study we aim to investigate this hypothesis by monitoring the friction evolution in several granular materials and bare surfaces. We compare velocity increases and decreases in both large and small velocity steps and investigate the role of out of shear material and shear zone thickness on brittle faulting.

### 3.2. EXPERIMENTAL METHODS

Experiments were conducted on three unique granular materials with varying grain sizes (Figure 3.2) and bare surfaces of Westerly granite. Quartz samples were obtained from the U.S. Silica Company, Rolla, MO. F110 quartz is medium-grained pure quartz sand and Min-U-Sil 40 is silt- to clay-sized pure quartz powder, and hereafter is referred to as fine-grained quartz. Caesar till is coarse grained sand obtained from the former Scioto Lobe of the Laurentide Ice Sheet, and composed of 35% quartz, 26% calcite 23% plagioclase, and 16% clay minerals, with clay mineral abundances of 35.3% smectite, 38.5% illite and 26.1% chlorite/kaolinite.*Rathbun et*
Westerly granite was precision ground to make square and parallel blocks to within 0.001” then sandblasted with 200 grit glass beads for roughness to promote stable sliding.

The use of F110 quartz provides us with the advantage of comparing our data with those of several other research groups and studies. This material is commonly used in laboratory studies of granular friction [e.g. Mair and Marone, 1999; Frye and Marone; Anthony and Marone, 2005; Samuelson et al., submitted] and is one the materials being used in the on-going SAFOD (San Andreas Fault Observatory at Depth) interlab comparison study [Lockner et al., 2009; http://www.geosc.psu.edu/~cjm/safod/]. Additionally, Westerly granite has a long history of use in experimental rock mechanics experiments dating back to the early 1900’s.

All experiments were conducted in a servohydraulic testing apparatus in the double-direct shear configuration. This configuration consists of two parallel shear zones with equal thicknesses and contact areas sandwiched between three steel blocks (Figure 3.3). Force was measured via BeCu load cells attached to each loading ram and displacement was measured external to the shear zone by Direct Current Displacement Transducers (DCDT). Details of the experimental apparatus can be found in [Mair and Marone, 1999; Karner and Marone, 2001; Frye and Marone, 2002; Rathbun et al., 2008]. Normal stress ranged from 1-40 MPa and was kept constant during shear. In all experiments the nominal contact area was kept constant at 10 cm x 10 cm. In our experiments the initial macroscopic shear zone thickness ranged from 1 cm in granular samples to 0 on bare surface experiments. Shearing
velocity was varied in a series of step changes ranging from 1 μm/s to 300 μm/s. Complete normal stress, layer thickness and shear velocity history for each experiment is presented in Table 3.1.

3.3. RESULTS

3.3.1 Velocity Stepping Experiments

We conducted a series of velocity step experiments on several materials under variable layer thickness and stress conditions to probe how localization affects RSF and to evaluate which law may be most applicable to our experiments. Velocity steps began at the initiation of the experiment. Initially, the coefficient of friction rapidly increases at low strain then becomes steady (Figure 3.3). We only consider velocity steps after friction has reached steady-state. Each velocity step results in a change in friction with velocity increases (decreases) leading to pronounced peaks (troughs) in friction followed by evolution back to steady-state sliding friction (Figure 3.3, inset). The difference between friction before the steps and the peak is associated with $a$, whereas $b$ is associated with the difference between the peak and the new steady state friction. The evolution from peak friction to the new steady state is controlled by $D_c$ which is the $e$-folding length from peak to steady state (Figure 3.3, inset). Due to the finite stiffness of the loading apparatus, the RSF parameters cannot be measured from a velocity step and must be modeled, accounting for the elastic loading stiffness (e.g. Equation 4, Figure 3.1).

To assess asymmetry or symmetry of frictional responses to velocity perturbations and to evaluate the RSF laws, we compare a series of velocity steps.
Figure 3.4a presents 10 consecutive velocity steps (five increases and five decreases) from an experiment conducted on F110 quartz at 25 MPa normal stress. Each step is plotted by lining up the point at which the velocity step occurs for easy comparison, with the velocity increases shown in black and decreases in red. The velocity decreases are inverted and shown as mirror images in green, e.g. Figure 3.1. Each velocity increase (decrease) shows remarkable reproducibility during the loadup portion before peak friction and the evolution to steady state. Similar to the synthetic steps presented in Figure 3.1, the velocity decreases have a higher apparent stiffness, as expected, as well as a smaller peak friction value. The measured $k$ in these experiments was 0.001 $\mu$m$^{-1}$, corresponding to the intermediate case of $k$ in Figure 3.1. During the decrease to steady sliding friction, the two types of steps match and evolve together beginning at ~25 $\mu$m, well before they reach steady state. Velocity steps for F110 quartz are near velocity neutral with some steps slightly velocity strengthening and some slightly weakening.

Caesar till (Figure 3.4b) displays similar behavior to F110 quartz. As with F110 quartz, velocity decreases and increases come to steady state at similar displacements, not with velocity decreases leading increases as predicted by the Dieterich law. The peak friction is larger for till than for quartz. The length of $D_c$ is also longer for till than for F110 quartz. In all cases velocity steps on till are velocity strengthening [e.g. Rathbun et al., 2008; Rathbun and Marone, 2010].

Velocity steps on fine-grained quartz (Figure 3.4c) again show reproducibility for velocity increases and decreases. Due to velocity weakening and stick-slip behavior at high normal stress, data for fine-grained quartz are only presented at 1
MPa normal stress. In experiments on the fine-grained quartz, velocity increases reach steady state sliding friction at ~10 μm of slip while velocity decreases reach steady sliding friction at ~25μm, contrary to the predictions of the Ruina and Dieterich laws.

For the three granular materials presented in Figure 3.4, the length of $D_c$ for steady sliding friction correlates with the average grain size. For the finest grained material, fine-grained quartz, steady friction is established by ~10 μm for velocity increases and ~ 25 μm for decreases. Increasing the average grain size to that of F110 quartz lengthens the distance to steady sliding friction to ~100 μm for both velocity increases and decreases. In the case of our largest grain sized material, glacial till, steady friction is not established until >200 μm.

Increasing the size of the velocity perturbation by an order of magnitude to 10-300 μm/s increases the size of the friction peak for both velocity increases and decreases and lengthens the displacement needed to reach a new steady friction level (Figure 3.5a). Again, each step is nearly indistinguishable for velocity decreases; however, the increases show minima ~75 μm after the velocity step. The step with the lowest friction minimum is the first step and they follow, in order, to the last step, which occurs at the largest displacement. This minimum in the friction is only observed for the velocity increases. The decreases show a smooth evolution to steady state, as in the smaller steps.

Till displays differences in behavior between increases and decreases for large steps (Figure 3.5b). Velocity decreases again show a smooth transition from steady state while increases display a break in the trend similar to Figure 3.5a. In both
Figures 3.5a and 3.5b, the first length scale associated with friction evolution in the velocity increases evolves at a shorter displacement for velocity increases than decreases. The second length scale increases the total displacement required to establish steady friction. As a result of the second length scale, velocity increases and decreases reach steady state at approximately the same displacement, but with different shapes. As with experiments at large normal stress, large steps on the fine-grained quartz display unstable behavior on velocity increases and therefore are not presented. The velocity decreases on fine-grained quartz display a smooth transition to steady sliding friction similar to both till and F110 quartz.

3.3.2 Dilation and Compaction During Velocity Steps

Each perturbation in velocity causes an associated change in layer thickness. Velocity increases dilate the layer and decreases compact the layer (e.g. Figure 3.6). Previous investigators have used the amount of dilation as a proxy for localization in a shearing layer [Marone and Kilgore, 1993]. We follow this work, comparing the relative amounts of dilation and compaction for the velocity steps in our experiments. Following standard procedure [i.e. Scott et al., 1994] we remove a linear trend of decreasing layer thickness produced from a geometric thinning in some shear apparatuses. Layer thickness change is presented for 10 consecutive steps in Figure 3.6. As with plots of friction evolution, the velocity step that perturbs layer thickness occurs at the first tick mark on the plot, and each step is offset to the same displacement for comparison. Figure 3.6a presents the change in layer thickness for representative steps on F110 quartz at 25 MPa with velocity steps between 10 and 30
μm/s. Dilation is given as a positive change in layer thickness and compaction as a negative change. As with the friction curves (Figures 3.4, 3.5) the compaction associated with velocity decreases is mirrored for comparison to velocity increases. After a velocity increase the layer dilates ~1 μm for an initially 10 mm thick sample, while after a decrease the layer compacts ~1.5 μm (Figure 3.6a). As with the friction curves, layer thickness trends are reproducible. Experiments on Caesar till have larger values for dilation/compaction and more variability between each step (Figure 3.6b). As with F110 quartz, the largest values of layer thickness change are associated with velocity decreases, but the two populations show overlap in layer thickness change. Experiments on fine-grained quartz display the largest disparity between velocity increases and decreases (Figure 3.6c). Decreases in velocity compact the layer ~2x as much as increases dilate the layer.

The change in layer thickness and the difference between compaction and dilation scales with grain size in our experiments. The largest grain size, till, has the largest dilation/compaction, followed by F110 quartz and then fine-grained quartz (Figure 3.6). The separation between the dilation and compaction in till is unclear; however, it appears that compaction is slightly larger than dilation (Figure 3.4b). In F110 quartz, compaction is generally larger than dilation with some overlap between the two, while in fine-grained quartz there is clear separation between dilation and compaction. [e.g. Samuelson et al. submitted]

Increasing the size of the velocity steps to 10 to 300 μm/s highlights the difference between velocity increases and decreases (Figure 3.7). Large velocity increases and decreases on F110 quartz show ~2x more compaction than dilation after
a velocity step (Figure 3.7a). The compaction increases from ~1.5 μm on small steps to 4 μm on large steps, while the dilation increases from 1 μm to ~2 μm for small and large steps, respectively. Comparing the increases vs. decreases in the large steps indicates that the rate of change of layer thickness is larger for velocity decreases. As with the dilation/compaction with small velocity steps, till displays less disparity than F110 quartz between velocity increases and decreases (Figure 3.7b), with greater changes in layer thickness in the larger grain sized till. As with the small steps till has a greater variability in the change in layer thickness than in F110 quartz.

3.3.3 Bare Surface Experiments

Two experiments on roughened bare granite surfaces show equal displacements needed to reach steady state for order-of-magnitude steps (Figure 3.8). Both bare-surface experiments show more variability than experiments on granular materials, but still yield reproducible results both between and within individual experiments. Velocity increases do show a tendency to produce unstable slip after, as evidenced by the trough in friction at ~ 8μm displacement after the velocity step (Figure 3.8a) Steps from experiment p2646 (Figure 3.8b) stably slide at 11 μm/s. This difference is likely caused by subtle differences in surface roughness during sand blasting.

Velocity increases and decreases required similar slip to reach steady state (Figure 3.8b). As predicted by the RSF laws coupled with elastic stiffness (Figure 3.1), velocity decreases reach peak friction at a smaller displacement than increases. Also, peak friction is smaller for decreases than increases. The evolution to steady
state then occurs with both increases and decreases reaching steady sliding friction at ~ 8 μm. As with the experiments on granular materials, the velocity decreases do not evolve to steady friction at smaller displacements. This evolution is most consistent with the Ruina slip law (Figure 3.1b). In contrast to the experiments on granular samples, the displacement needed to reach steady state is equal for all cases in bare surface experiments.

3.3.4 Normal Stress Oscillations

To evaluate the role of so-called spectator regions within shear zones [e.g. Mair and Hazzard, 2007] we conducted a series of experiments in which velocity steps were imposed on the sample, and then normal stress was oscillated to ensure that the layer was fully compacted, followed by another series of velocity steps. In one experiment normal stress was oscillated in 64 cycles between 25 and 15 MPa, with all shear conducted at 25 MPa (Figure 3.9). Figure 3.9a presents the complete history for experiment p2648. Shear occurs until ~13mm at 25 MPa (Figure 3.9a) and then shear stress is removed. During the normal stress oscillations (Figure 3.9c) the layer compacted from ~7.65 mm to ~7.59mm corresponding to a porosity loss of 0.8 porosity units (Figure 3.9b). In another experiment stress was increased in a series of 8 cycles between 25 and 35 MPa, corresponding to a thickness change of 30 μm or 0.4 porosity units with all velocity steps conducted at 25 MPa normal stress before and after oscillations.

Velocity steps before and after stress oscillations are presented in Figure 3.10. The oscillations reduce both the peak friction and the distance required to evolve back
to steady state (Figure 3.10a, b). Figure 3.10b presents forward models overlying the
two populations of velocity steps. RSF parameters are \( a = 0.0073, b = 0.00625 \) and \( D_c = 36 \ \mu m \) before oscillations and \( a = 0.0075, b = 0.007 \) and \( D_c = 16 \ \mu m \) after oscillations using the Ruina law. We choose to only use one state variable in the
models that are shown even though there is considerable misfit (Figure 3.9b). This
highlights the effects of the normal stress oscillations changing the behavior of the
shearing layer. In both experiments the peak friction associated with \( a \) is smaller after
the stress oscillations. Oscillations also decrease the size of the layer dilation. Before
normal stress was oscillated the layer dilated \( \sim 1.5 \ \mu m \), and after the oscillations \( \sim 0.75 \ \mu m \). This difference can be interpreted as a decrease in the thickness of the active
shear zone, or localization.

### 3.4. DISCUSSION

#### 3.4.1 Which Law?

The Ruina slip law best matches our experimental data. The Dieterich law
predicts that velocity decreases should reach steady sliding friction at a smaller
displacement than increases. Our experiments on three different granular materials
and bare granite surfaces produces result contrary to this prediction of the Dieterich
law. Conversely, we observe that velocity increases and decreases reach steady state
at similar displacements (Figure 3.4a, 3.8), or that velocity decreases reach steady
state at smaller displacements than increases (Figure 3.4c) in experiments with small
velocity steps. Increasing the size of the velocity steps from 10-fold to a 30-fold
causes velocity increases to display two-state behavior while decreases still only have
one length scale to reach steady state (Figure 3.5). The first length scale on the large increases evolves friction at a much smaller slip than velocity decreases. The second length scale then evolves friction at a similar displacement as decreases.

These data agree with the hypothesis of Rathbun and Marone [2010], who showed that for experiments at 1 MPa normal stress, a longer critical slip distance was required to fit their data with velocity decreases when using the Dieterich law. This led to the assertion that the Ruina law best described their experiments. This is in contrast to the healing experiments of Beeler et al. [1996], which showed that the Dieterich law best matched their data. We observe behavior consistent with neither the Ruina nor Dieterich laws in that velocity increases can evolve to steady sliding friction at smaller displacement than velocity decreases, i.e. Figure 3.4c.

Several other attempts have been made to update the evolution laws. Neither of the commonly used laws has been adequate in describing all laboratory data and different variables that may control the natural system. In the standard RSF equation no term is included for variations in normal stress; however, work by Linker and Dieterich [1992] has extended Equation 1 to include normal stress. Chester and Higgs [1992] incorporated temperature in the RSF equations. Perrin et al. [1995] extended the evolution laws to include symmetry between velocity increases and decreases similar to the Ruina law, but with time-dependent aging like the Dieterich law. A composite law was proposed by Kato and Tullis [2001] to explain the experiments of Beeler et al. [1994]. The RSF laws also do not include chemical effects that have been shown to interact with frictional interactions [e.g. Bos et al., 2000; Niemeijer et al., 2008]. In our study we only attempt to evaluate the two most
commonly used laws that form the basis for most models of earthquake rupture [e.g. Heaton, 1990; Beeler and Tullis, 1996; Ampuero et al., 2002; Lapusta and Rice, 2003; Ziv and Rubin, 2003] not test all of the alternative forms. 

Sleep [2005] attempts to attach a physical meaning to the RSF friction laws, contrasting the physical basis for each of the two common laws. They concentrate on healing when \( V=0 \), rather than sliding. Sleep [2005] shows that the slip law arises from exponential creep at contacts and scales with contact size, while the aging law from creep for both shear and compaction at the subgranular scale. They attempt to place bounds on the applicability of each law with the slip law occurring at low humidity and the aging law at high humidity [i.e. Frye and Marone, 2002]. We find that the slip law better approximates our data in experiments conducted at room temperature and humidity.

3.4.2 Layer Controls on Friction Parameters

We propose that differences between velocity increases and decreases are caused by changes in the shearing layer. Both suites of experiments, velocity step comparison of increases and decreases and normal stress oscillations show that grain-to-grain interactions and the characteristics of the localized shear zone control the rate and state friction parameters.

For both large and small velocity steps an inequality in the dilation verses compaction exists. During a velocity decrease, shear slows within the active, localized zone. Both the localized zone and the passive, out-of-shear zone feel the same normal traction. Because an equal traction occurs on both the localized and
passive zone, the entire layer is available to compact, with the possibility of a
difference in compaction between the active and passive zones of the layer. During a
velocity increase, grains in the active zone must move around each other to allow for
shear, while the out-of-shear material is not affected by the change in driving
velocity. This movement results in the macroscopic dilation observed in experiments.
With the case of a localized shear zone, we expect that dilation will be smaller than
compaction because only the localized zone can dilate but the entire layer can
compact. For the factor of three steps the layer compacts slightly more than it dilates
after a velocity step (Figure 3.6). When the step size is increased to 30x the separation
between dilation and compaction widens (Figure 3.7). The difference between
dilation and compaction suggests differences in the micromechanics of the layer
during each of the directional changes in velocity. During dilation, only the active
portion of the layer needs to dilate as a result of the velocity change. This is in
contrast to the velocity decreases and compaction when the entire layer is available to
change in width.

The velocity stepping experiments show two-state behavior with dilatancy
controlling one length scale and the evolution of frictional contacts controlling the
other. In the case of comparing velocity increases and decreases, the combination of
dilatancy and contact evolution work to produce an asymmetry in the frictional
response. This asymmetry is not predicted by either the Ruina slip law, which
predicts symmetry, or the Dieterich aging law, which predicts asymmetry opposite
that of our experiments.
The idea that layer effects obscure the traditional RSF laws is supported by experiments in which normal stress was oscillated. Performing these normal stress oscillations reduced porosity and compacted our layer. After the oscillations the dilation decreased and the critical slip distance required to reestablish steady sliding friction shortened. Both the displacement needed to establish steady friction (Figure 3.10c) and the dilation associated with a velocity step (Figure 3.10b) decreased by approximately a factor of two after normal stress oscillations.

The decrease in both $D_c$ and $a$ is consistent with a decreasing shear zone thickness after the stress oscillations. A thicker active shear zone and more distributed deformation promotes a larger $a$ as more bonds need to be broken after the velocity perturbation. The evolution of the $D_c$ also points to an increase to the degree of localization after the stress oscillations. Marone and Kilgore [1993] showed that $D_c$ can be correlated to shear zone thickness. In our experiments $D_c$ decreases by $\sim 2x$ as a result of compacting the macroscopic layer. This reduction in $D_c$ suggests that the shear zone becomes more localized as the result of the normal stress vibrations. This enhanced localization decreases both the direct effect and critical slip distance.

The slip required to reach steady driving friction (i.e. Figure 3.4) can be understood in terms of shear zone thickness. In granular materials shear localizes into discrete zones several to $\sim 20$ particles thick [e.g. Muhlhaus and Vardoulakis, 1987; Tordesillas et al., 2005; Rathbun and Marone, 2010]. The length of $D_c$ has been shown to correlate with the thickness of the active shear zone of a material [Marone and Kilgore, 1993]. We propose a model in which slight changes in the active shear zone overrides the expectations of the evolution laws.
During a velocity increase the layer initially dilates (Figure 3.6, 3.7). This dilation slightly decreases the contact area, $A_r$, between particles in the granular shear zone. Each particle contact is likely velocity weakening [Marone et al., 1990], which promotes shear to further localize. Initial dilatancy leads to a localized system and a more rapid evolution to steady friction. After a velocity decrease the layer compacts (Figure 3.6, 3.7) increasing the contact area of particles in the active zone leading to grain strengthening and locking. The growth of $A_r$ between particles dominates the system and the central portion of the sample is no longer the weakest area. Compaction increases the sliding friction of the system by increasing $A_r$. Because contacts need to be evolved, $D_c$ lengthens in the velocity decrease. In granite block experiments, with no variation in shear zone thickness, we see no observable dilation or compaction and the expected RSF response occurs. Dilation leads to a narrowing of an active zone by the reduction of $A_r$ and localization while compaction leads to thickening of the shear zone by increasing $A_r$, between grains and widening the shear zone.

### 3.4.3 Two-State Behavior

The large velocity increases show two distinct length scales to reach steady state (Figure 3.5). Velocity increases have a pronounced, sharp, frictional peak that decays toward a steady state, and then incorporates another length scale for both F110 quartz and till. Two-state behavior is not uncommon and has been noted by many investigators [e.g. Tullis and Weeks, 1986; Cox, 1990; Marone and Cox, 1994; Blanpied et al., 1998]. Cox [1990] argued that two-state behavior was caused by a
longer $D_c$ related to structure in the gouge zone while the shorter $D_c$ was related to the evolution of surface properties. Marone and Cox [1994] conducted experiments on bare surfaces of gabbro with varying surface roughness. They found that the second $D_c$ disappeared with increasing displacement. This led Marone and Cox [1994] to conclude that their $D_{c2}$ was a surface effect and that $D_{c1}$ was a property of the gouge. They concluded that $D_c$ in granular experiments could be thought of as the accumulation of several $D_c$ from grain-to-grain interaction [i.e. Marone and Kilgore, 1993].

In our experiments we observe a difference in the compaction/dilation of the layer associated with velocity steps, which we argue is the result of localization. Caesar till has been shown to localize shear into a finite boundary-parallel zone [Rathbun and Marone, 2010], whereas F110 quartz is well known to localize shear onto Y and R shears [Mair and Marone, 2000]. We infer that the two-state behavior observed is the result of changes in the micromechanics of the localized shear zone. During velocity increases, the first length scale is associated with grain-to-grain contacts and frictional evolution and the second with a length to dilate the localized shear zone. It seems likely that a length scale to dilate the layer is present in steps of all sizes; however, that $D_c$ is only observed when the step size increases to a large enough magnitude.

The large velocity increases also highlight the asymmetry in friction evolution. In contrast to the break in the evolution to steady state for velocity increases, velocity decreases show a smooth transition to steady state. This behavior agrees with changes in the localized zone, as the result of dilation, controlling the
friction evolution. The first length scale for state evolution suggests a shorter
displacement to steady state for the velocity increases vs. the decreases until the
second state evolution dominates. This behavior matches with the Dieterich
formulation of the state evolution law until the second length scale dominates.

3.4.4 Implications for the Stability of Fault Zones

Fault zone stability is often thought of in terms of critical stiffness. The fault
is unstable when a critical stiffness exceeds the loading stiffness [Rice and Ruina,
1983],

\[ k < k_c = \frac{(b-a)\sigma_n}{D_c} \left[ 1 + \frac{mV^2}{\sigma_s aD_c} \right] \] (5)

where \( a, b \) and \( D_c \) are the RSF parameters, \( m \) is the mass per unit area and \( V \) is velocity. The first term is sufficiently large that the second-order term is often ignored, yielding:

\[ k < k_c = \frac{(b-a)\sigma_n}{D_c} \] (6)

In this formulation the RSF parameters \( a, b, D_c \) and the normal stress, \( \sigma_n \), define a critical stiffness, \( k_c \). When the critical stiffness exceeds the stiffness of the laboratory apparatus and sample or the stiffness of the crustal rocks in natural systems, conditions are sufficient for earthquakes to occur.

In experiments where shear is stopped and normal stress is oscillated, both \( a \) and \( D_c \) decrease after the vibrations leading to a more unstable fault zone. It is a necessary, but insufficient condition that the \((b-a)\) term in Equation 5 is positive, the velocity weakening condition, for an earthquake to occur. A decrease in the peak
friction as shown in Figure 3.10, yields a larger term for \((b-a)\) and could transition a material from velocity strengthening to velocity weakening.

The decrease in \(D_c\) promotes seismic behavior in the fault zone. It should be assumed that seismogenic fault zones are already velocity weakening, satisfying the necessary conditions of Equation 5. The reduction of \(D_c\) promotes unstable slip and the possibility of the occurrence of an earthquake by increasing the critical stiffness, \(k_c\), of the fault zone. It is possible that this grain rearrangement may be a mechanism for earthquake triggering.

3.5. CONCLUSIONS

We find that in velocity stepping experiments on three unique granular materials that the Ruina slip law better approximates our data than the Dieterich aging law. We show that on velocity-strengthening glacial till, weakening fine-grained quartz and F110 quartz which transitions from strengthening to weakening that an asymmetry between velocity increases and decreases can occur. Experiments on bare surfaces of Westerly granite blocks produce a symmetric or near symmetric response for both velocity increases and decreases. The asymmetry we observe is not predicted by either of the commonly used laws. We propose a new conceptual model based on the micromechanics of the granular shear zone to explain this asymmetry. An additional suite of experiments was conducted in which normal stress was oscillated in between two series of standard velocity steps. We find that both the critical slip distance and the direct effect decrease as a result of these oscillations supporting a model in which shear localization in an active zone controls the frictional response.
Shear localization also works to produce two distinct length scales for frictional evolution, causing two-state behavior in our experiments. Localization produces a smaller critical slip distance, which enhances the likelihood of seismic slip.

3.6. ACKNOWLEDGEMENTS

Grain size characterization was conducted at the Materials Characterization Lab (MCL) at Penn State University. This work benefited from comments by Bryan Kaproth.

REFERENCES


Chester, F. M., and J. S. Chester (1998), Ultracataclasite structure and friction processes of the Punchbowl fault, San Andreas system, California, Tectonophysics. 295, 199-221.


Heaton, T.H (1990), Evidence for and implications of self-healing pulses of slip in earthquake rupture, phys. Earth Planet Inter. 64, 1-20.


<table>
<thead>
<tr>
<th>Experiment</th>
<th>Condition</th>
<th>Material</th>
<th>Layer Thickness $\mu$m</th>
<th>Normal Stress MPa</th>
</tr>
</thead>
<tbody>
<tr>
<td>p1345</td>
<td>Sat</td>
<td>C. Till</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p1507</td>
<td>Sat</td>
<td>C. Till</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p1968</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p1969</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p1970</td>
<td>Dry</td>
<td>F.G. Quartz</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p1971</td>
<td>Sat</td>
<td>F110 Quartz</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2064</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2065</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>3</td>
<td>1</td>
</tr>
<tr>
<td>p2410</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>10</td>
<td>25</td>
</tr>
<tr>
<td>p2411</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>10</td>
<td>25</td>
</tr>
<tr>
<td>p2412</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2413</td>
<td>Dry</td>
<td>F.G. Quartz</td>
<td>10</td>
<td>25</td>
</tr>
<tr>
<td>p2414</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2415</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>10</td>
<td>25</td>
</tr>
<tr>
<td>p2416</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2443</td>
<td>Dry</td>
<td>F.G. Quartz</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2444</td>
<td>Dry</td>
<td>F.G. Quartz</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2445</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2447</td>
<td>Dry</td>
<td>C. Till</td>
<td>10</td>
<td>25</td>
</tr>
<tr>
<td>p2636</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>10</td>
<td>25-35 8 cycles</td>
</tr>
<tr>
<td>p2637</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>10</td>
<td>35</td>
</tr>
<tr>
<td>p2638</td>
<td>Dry</td>
<td>Granite Blocks</td>
<td>0</td>
<td>5</td>
</tr>
<tr>
<td>p2645</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>7</td>
<td>25</td>
</tr>
<tr>
<td>p2646</td>
<td>Dry</td>
<td>Granite Blocks</td>
<td>0</td>
<td>5</td>
</tr>
<tr>
<td>p2647</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>3</td>
<td>25</td>
</tr>
<tr>
<td>p2648</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>10</td>
<td>25-15 64 cycles</td>
</tr>
<tr>
<td>p2649</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>10</td>
<td>25</td>
</tr>
<tr>
<td>p2650</td>
<td>Dry</td>
<td>F110 Quartz</td>
<td>10</td>
<td>25</td>
</tr>
</tbody>
</table>

Table 3.1 Experiment table.
Figure 3.1. Synthetic velocity steps for the Dieterich law (a) and Ruina law (b). Four velocity increases (black) and four decreases (red) occur at displacement equal to 0, between 10-30 μm/s. Velocity decreases are flipped and presented as mirror images for comparison to increases (green). Rate and state parameters are $a = b = 0.007$, and $D_c = 30 \mu m$. Elastic loading stiffnesses are 0.004 $\mu m^{-1}$, 0.0005 $\mu m^{-1}$, the stiffness of our experiments, 0.001 $\mu m^{-1}$ and a theoretical infinitely stiff case. The infinitely stiff case is dashed to highlight the identical behavior pre-peak of increases and decreases and evolution in (b).
Figure 3.2. Grain size distribution of the three granular materials used in this study. Grain size is determined via the laser absorption method for F110 and fine-grained quartz samples. Caesar till is sieved to 0.064mm then measured via laser absorption.
Figure 3.3. Friction-displacement curve for an entire experiment. Normal stress, $\sigma_n$, is held constant horizontally (bottom right) with a constant shear velocity applied at the top of the three block arrangement. Shearing velocity was stepped between values indicated on the top of the figure. Velocity steps begin at the initiation of shear with each step last 450 µm. Rate and state friction parameters are given in the inset.
Figure 3.4. Velocity increases (black) decreases (red), and mirrored decreases (green) for F110 quartz (a), Caesar till (b) and fine-grained quartz (c). For each panel, n corresponds to the number of steps. In all cases the velocity steps were preformed between 10 and 30 μm/s with (a) and (b) at 25 MPa normal stress and (c) at 1 MPa. All steps are consecutive and offset such that the displacement and friction equal 0 at the point in which the velocity step occurs. Friction is plotted as the change from steady state sliding friction prior to the velocity step.
Figure 3.5. Ten consecutive velocity steps between 10-300 μm/s for each of F110 quartz (a) and Caesar till (b). In both panels, velocity increases are shown in black, decreases in red and decreases are mirrored in green. In both experiments the normal stress was 25 MPa. For each panel, \( n \) corresponds to the number of steps.
Figure 3.6. Change in layer thickness as a result of a velocity increase (black) or decrease (red) with the decreases mirrored (green). Upon a velocity increase the layer dilates, a positive change in thickness, and after a decrease the layer compacts. The legend in panel (a) applies to all three panels, with 5 consecutive steps of increases and decreases presented with all steps between 10-30 μm. (a) F110 quartz at 25 MPa normal stress, (b) Caesar till at 25 MPa (c) fine-grained quartz at 1 MPa.
Figure 3.7. Change in layer thickness for velocity steps between 10-300 μm/s in F110 quartz (a) and Caesar till (b) with the layer thickness changes corresponding to the velocity steps presented in Figure 3.5.
Figure 3.8. Comparison of order of magnitude velocity increases and decreases for two experiments on bare Westerly granite surfaces. Normal stress was 5 MPa in both experiments. In experiment p2638 (a) the drop in friction at ~10 µm displacement is related to unstable behavior. In both panels $n$ corresponds to the number of velocity steps.
Figure 3.9. Stress oscillations in experiment p2648. (a) Velocity steps are conducted at a constant normal stress of 25 MPa before and after normal stress oscillations. Shear stress is removed after ~13 mm of displacement and normal stress is oscillated (b) at a constant rate between 25 and 15 MPa for 64 cycles. Normal stress is then held constant during velocity steps. (c) Layer thickness change as a result of layer compaction during the stress oscillations.
Figure 3.10. Velocity steps before and after normal stress oscillations. Velocity is increased from 10 to 30 \( \mu \text{m/s} \), (a) Five steps \((n)\) immediately preceding (thin black lines) following (thin blue) the oscillations. Oscillations work to decrease the peak friction and decrease the length to a new steady sliding friction. (b) The velocity steps presented in (a) along with forward models for the RSF parameters (bold lines) using one state variable. Parameters for the reference models are \(a = 0.0073, b = 0.00625 (a-b = 0.00105)\) and \(D_c = 36 \mu \text{m}\) and \(a = 0.0075, b = 0.007 (a-b = 0.0005)\) and \(D_c = 16 \mu \text{m}\) for before and after normal stress oscillations, respectively. (c) Change in layer thickness associated with the velocity steps presented in (a, b).
Chapter 4. A NEW MECHANISM FOR SLOW-SLIP

Andrew P. Rathbun\textsuperscript{1,2}

Chris Marone\textsuperscript{1,2}

\textsuperscript{1}Department of Geosciences, Penn State University, University Park, PA 16802

\textsuperscript{2}Penn State Sediment and Rock Mechanics Laboratory
ABSTRACT

Slow-slip and earthquakes have been documented in both strike-slip and subduction zone faults around the world [Rogers & Dragert, 2003; Obara, 2002; Shelly et al. 2010; Delahaye, 2008], yet the understanding of the slow-slip mechanisms is incomplete. Elevated pore-fluid conditions and dilatancy coupled with rate and state friction are usually invoked to explain slow-slip [Rubin, 2008; Segall et al. in press] and the arrest of fast glacial slip [Moore & Iverson, 2002]. We show that slow-slip can be produced in laboratory shear experiments with a duration ranging from 1s to 100’s s in configurations similar to laboratory earthquakes in the absence of pore fluid. We compare slip, stress drop and seismograms of these events to typical laboratory stick-slip with slip duration < 0.001 s. We find that slip velocity scales with stress drop for many materials and that slip roughly scales with the rate and state friction parameter \((a-b)\); for both positive values, which are expected to slide stably and negative values, which may stick-slip. We propose an alternative model for slow-slip where accelerating creep rupture leads to slip rather than rate and state friction and fluid pressurization. Our results in the creep configuration have unique implications for glacial sliding [Moore & Iverson, 2002], landslides [Voight, 1989], earthquake triggering, earthquake prediction [Voight, 1989] and as an explanation of Omari’s Law [Main, 2000] versus traditional laboratory sliding experiments.
Slow-slip and non-volcanic tremor have been documented in Cascadia [Rogers & Dragert, 2003], Nankai [Obara, 2002], New Zealand [Delahaye et al. 2009], and the San Andreas Fault [Shelly et al. 2010]. Several modes of non-impulsive slip events have been documented (we will now refer to the family of these events as slow-slip) with magnitude scaling with duration [Ide et al. 2007]. Slow-slip events can last from seconds to years [Ide et al, 2007] Fig. 4.1. While these events represent up to ~M8, no theory has been able to explain all aspects of their occurrence. Proposed models for slow-slip typically invoke fluid pressurization of the fault zone coupled with rate and state friction [Rubin, 2008; Segall et al 2010]; however, modeling is yet to capture the rupture velocity in slow-slip.

The current understanding of brittle faulting hinges on the rate and state friction laws [Deiterich 1978; 1979; Ruina 1981] in which a decrease in friction is required with increased slip velocity, also known as a velocity-weakening. Stick-slip (Fig. 4.2) may occur when both the frictional strength at the interface is exceeded by the shear stress and the critical stiffness of the shear zone \((k_c)\) exceeds the stiffness \((k)\) of the loading system [Ruina, 1981, Gu et al. 1984].

\[
k < k_c = \frac{-(a-b)\sigma_n}{D_c} \tag{1}
\]

The critical stiffness is defined by the velocity dependence of friction \((a-b)\), normal stress \(\sigma_n\) and the critical slip distance \(D_c\) of the frictional contact area. When an increase of driving velocity results in an increase in sliding friction \((a-b)\) is positive) the contact is velocity-strengthening, when increased velocity results in lower friction \((a-b)\) is negative) the contact is velocity-weakening and the nucleation of slip is possible [Scholz, 1998; Marone, 1998].
We conduct standard experiments where shear rate is controlled to produce stick-slip (Fig. 4.2) as well as constant loading stress (creep) experiments to produce slow-slip. While creep experiments are common in soil mechanics and rock fracture they typically involve only compression with no shear component. The slow-slip we report on was created in the creep configuration in double-direct shear (Fig. 4.2 inset). During the slip event (Fig. 4.3) displacement rapidly accelerates in tertiary creep, then transitions back to primary and secondary creep, with the stress drop occurring during the tertiary creep phase. During the slip event the material lacks sufficient strength to support the applied shear stress, resulting in acceleration. After some time, stored strain energy is released and stress begins to accumulate towards the imposed value. As the transition from stress drop to stress accumulation occurs, the resulting change in velocity produces a switch from a positive acceleration to negative acceleration.

We use four materials covering the spectrum of rate and state behavior. Till is strongly velocity-strengthening [Rathbun et al 2008], F110 and fine-grained pure quartz transition from velocity-strengthening at low strain to velocity-weakening at high strain. Fine-grained quartz displayed stick-slip at normal stress > 5 MPa and spherical glass beads stick-slip at all stress and strain conditions, which indicates extreme velocity-weakening. Despite a wide range of friction behavior, all the materials display the same form of slip and stress drop in slow-slip events (Fig. 4.3) an unexpected result from rate and state friction.

The duration of the slip event scales with \((a-b)\) for the materials (Fig. 4.3). Events in till take 100’s s for stress to drop and recover, F110 quartz events 10’s s, seconds in fine-grained quartz and ~1s in glass beads. The slow-slip of glass beads is
accompanied by audible stress drops typical of stick-slips. Traditional stick-slips in beads (presented in Fig. 4.2) occur faster than our highest recording rate of 1000Hz indicating a slip duration ~ 1 ks.

An accelerometer records an acceleration seismic record (Fig. 4.4). Our slow-slip events are more impulsive with larger amplitude and shorter duration than stick-slip. We observe a ~100x greater amplitude for the stick-slip than slow-slip event in the acceleration seismogram. In the stick-slip a short ~0.001 s precursory release is observed followed by most of the seismic release over another ~0.001 s, and then another ~0.001 s of low amplitude seismic displacement. The slow-slip event is correlated with the shear stress record by aligning the first motion in both records. Two precursory stress drops are accompanied by seismic energy at t = 0 and t=0.1 s. As the amplitude of the seismic record approaches the peak value, the shear stress release accelerates. The transition to recovery begins just after the peak seismic amplitude. The envelope of the seismic record decreases with the cessation of the event when shear stress is recovered. The entire event lasts ~1 s with stress evolution lasting longer than the release of seismic energy.

Stress drop size controls the slip velocity over several orders of magnitude in our slow events (Fig. 4.5a). For approximately equal stress drops slow-slip events achieve a slip velocity up to 2.5 mm/s, while stick-slips slip up to 45 mm/s when a duration of 0.001s is assumed. These velocities are in line with other published durations and velocities in laboratory experiments [Okubo & Dieterich, 1981]. All materials follow the same trend, indicating that this process is controlled by an alternative factor than material properties. For stick-slips, the stress drop defines the
displacement of an event while slow-slip events are more scattered (Fig. 4.5b). We find that the stored strain energy controls the slip velocity. When a large stress drop occurs, the strength of the slipping region is far from the imposed stress of the loading system, which results in faster motion.

Rock fracture has been described in terms of constant stress rupture [Reches & Lockner, 1994; Lockner, 1998; Main 2000]. In these models, the components of creep are described as the interplay of distributed tensional microcracking in primary creep and the coalescence of microcracks into a plane during tertiary creep with the two processes approximately balanced during secondary creep. In a granular material this same process can be thought of as the failure of grain-to-grain contacts. During shear, bonds break and form between particles. During primary creep bond formation dominates while in tertiary creep bond destruction dominates [Mitchell & Soga, 2005]. Locker, 1998 presents a model for rock failure in creep similar in form as rate and state friction, albeit with no form of healing, which is required for repeated events. We propose that slow-slip events in the laboratory can be thought of as the release of elastic strain energy stored at frictionally governed contacts. These contacts fail throughout the layer yielding primary and secondary creep until they coalesce at a localized area leading to tertiary creep. Once tertiary creep relieves the critically stressed contacts, slip transitions back to the stable phase.

We show that in laboratory experiments slow-slip and stick-slip can be propagated spontaneously under simple conditions. As with natural events, slip duration for our slow-slip lasts several orders of magnitude longer and contains more relative energy at low frequency. These events occur in both velocity-strengthening
and weakening materials, which is inconsistent of the predictions of earthquake propagation from rate and state friction and dilatant hardening. Events in glacial till occur in both the presence and absence of pore fluid and all other experiments are conducted with no fluids.

**Methods**

Experiments are conducted in the double-direct shear configuration with two granular shear layers sandwiched between three grooved steel forcing blocks (Fig. 4.2 inset). Normal and shear tractions are applied via two servo-hydraulic rams with normal stress held constant (horizontal ram) and shear velocity controlled to produce stick-slip or shear stress controlled to produce slow-slip. In all cases the nominal frictional contact area is kept constant at 10cm x 10cm with initial layer thicknesses of 1cm, normal stress is held constant at 1 MPa to prevent grain breaking and associated effects. All experiments on glass beads and quartz are conducted at room humidity while experiments on till were conducted in either drained saturated conditions or at room humidity. A Brüel & Kjaer type 4393 accelerometer is attached to side of the center block (in and out of the paper Fig. 4.2 inset) in all experiments except for p2774 in which it is placed on the top of one side block.

Forces are measured via BeCu load cells to a resolution better than 0.1 kN, and displacement is measured on each loading ram to a precision better than 0.1 μm. Stresses and displacements are recorded at 10 kHz and averaged to a 10-1000Hz depending on the experiment and slip duration. The accelerometer has a frequency range of 0.1Hz to 16.5 kHz and is recorded at 1 MHz.
Creep experiments were conducted by first running a constant shear displacement rate to measure the frictional strength and subsequently the machine was switched to a constant shear force [Rathbun et al. 2008, Rathbun and Marone 2010]. Shear stress was then stepped until reaching ~90-105% of the frictional strength. Tertiary, accelerating creep occurs at either the completion of the stress step or during the stress hold. In experiments on glass beads we often obtain more than 15 slip events at the same shear stress. Caesar till is obtained from the former Scioto lobe of the Lauerentide ice sheet, Columbus Ohio, F110 and Min-U-Sil 40 pure quartz samples were obtained from the US Silica Company, Ottawa, Illinois and soda-lime glass beads from Mo-Sci Corporation, Rolla Missouri. Grain size distributions are presented as supplementary Figure S1.

Acknowledgments

We thank Matt Knuth for assistance with recording and Mike Cleveland with help processing seismic signals. Grain size measurements were conducted in the Materials Characterization Lab (MCL) at Penn State University. This paper benefited from comments by Luke Zoet.

References

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Material</th>
<th>Layer Thickness (mm)</th>
<th>Norm. Stress (MPa)</th>
</tr>
</thead>
<tbody>
<tr>
<td>p757</td>
<td>C. Till (Dry)</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p758</td>
<td>C. Till (Saturated)</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p761</td>
<td>C. Till (Dry)</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2748</td>
<td>Fine-grained quartz</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2749</td>
<td>Beads</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2758</td>
<td>Beads</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2771</td>
<td>Beads</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2772</td>
<td>Beads</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2773</td>
<td>Beads</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2774</td>
<td>Beads</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2823</td>
<td>F110</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2824</td>
<td>F110</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>p2910</td>
<td>Fine-grained quartz</td>
<td>10</td>
<td>1</td>
</tr>
</tbody>
</table>

**Table 4.1** Experiment table.
Figure 1. Characteristic slip duration for natural events and laboratory earthquakes, after Ide et al. [2007]. Low frequency earthquakes (LFE), Very low frequency earthquakes (VLF), slow-slip events (SSE) episodic tremor (ETS) and other natural events are taken from Ide et al. [2007], laboratory stick-slip and slow-slip are from this study.
Figure 2. Typical laboratory stick-slip. Normal stress ($\sigma_n$) is held constant and shear velocity in controlled. Stress drop (presented as sliding friction) occurs in $<0.001$ s during which time rapid slip occurs. The acceleration seismogram for this event is presented in Fig. 4.4b. The imposed slip rate is 20 $\mu$m/s. (inset) Double-direct shear geometry. Two granular layers are sandwiched between three steel forcing blocks. The accelerometer is placed either on the side or the center block or top of the side block, indicated by triangles.
Figure 3. Friction ($\mu$) and fault slip, presented as both displacement and strain for four materials ranging from velocity-strengthening glacial till (a), F110 quartz which transitions from velocity strengthening to weakening (b), fine-grained quartz which transitions from strengthening to weakening and stick-slips at normal stress $> 5$ MPa (c), glass beads which stick-slip in constant shear velocity experiments (d). During the stress drop, slip accelerates then slows during the recovery of shear stress. Slip duration scales with the rate and state parameter, $a$-$b$. Time, displacement and shear strain are set to zero at the initiation of the stress drop for comparison between materials.
Figure 4. Acceleration seismograms for the slow-slip (a) presented in Fig. 4.3d and stick-slip from Fig. 4.2 (b) from experiment p2773. Both are aligned with the first seismic energy appearing at $t = 0$ s. Super imposed on the slow-slip signal is the friction-time function. Both slow-slip and stick-slip display the same shape with different time scales for energy release. Small and precursory events are observed in the friction and acceleration records.
**Figure 5.** (a) Maximum slip velocity and stress drop (presented as friction) define a linear relationship for slow-slip events. Velocity is calculated as a running average of time and displacement and likely does not capture the largest values above 2 mm/s. The inset shows the same data on log scales for comparison with stick-slips. Velocity is calculated on stick-slips by assuming a time of 0.001s and using the measured slip. (b) Stick-slips define a line of slip against stress drop. Fiducial lines are fit through the average value for each material with an intercept equal to zero.
S1. Grain size distribution. Till is sieved to <0.5 mm and then the laser absorption method is used. To preserve boundary conditions and ensure shear within the layer all grains > 1 mm are discarded. All other materials are analyzed by laser absorption.
**S2.** Complete history of experiment p2773. Initially shear displacement rate is controlled at 20 μm/s and then the experiment is switched to shear stress control (indicated by vertical line at t = 900 s). The thick vertical lines in shear stress correspond to stress drops as stick-slip (n = 22) in rate control and slow-slip (n = 27) is stress controlled portions of the experiment. Each slow-slip event is accompanied by displacement of the fault and is seen as the step increase in displacement. The step increases in shear stress during the stress control portion of the experiment are increases in shear stress to near the frictional strength. The two vertical arrows indicate the stick-slip shown as Fig. 4.2, 4.4a and the slow-slip shown in Fig. 4.3d, 4.4b.
Chapter 5: Permeability of the San Andreas Fault at depth

Andrew P. Rathbun\textsuperscript{1, 2}
Insun Song\textsuperscript{3, 2, 3}
Demian M. Saffer\textsuperscript{1, 2}
Chris Marone\textsuperscript{1, 2}

\textsuperscript{1}Department of Geosciences, Penn State University, University Park, PA 16802
\textsuperscript{2}Penn State Sediment and Rock Mechanics Laboratory
\textsuperscript{3}The Korea Institute of Geoscience and Mineral Resources, Daejeon 305-350, Korea.
ABSTRACT
Quantifying fault-rock permeability is important toward understanding both the regional hydrologic behavior of fault zones, and poro-elastic processes that affect fault mechanics by mediating effective stress. We conducted experiments on fault core from ~2.7 km depth from the San Andreas Fault (SAF) collected during the SAFOD drilling project. Experiments were conducted on the Central Deformation Zone (CDZ) which accounts for ~90% of the casing deformation measured between drilling phases. The CDZ is 2.6 m thick with a matrix grain size < 10 μm and ~5% vol. clasts. Permeability experiments were conducted as constant rate of strain (CRS) tests in uniaxial stress conditions or flow-though and constant pressure differential experiments in isostatic conditions on sub-cores perpendicular to the CDZ. We found that the permeability, $k$, of the CDZ decreases rapidly to ~$10^{-19}$ m$^2$ at an effective stress of 20 MPa. At an effective stress of 75 MPa, $k$ ranges from $8x10^{-21}$ to $4x10^{-20}$ m$^2$. Our results are consistent with published geochemical data from SAFOD mud gas samples and inferred pore pressures during drilling [Zoback et al., 2010], which together suggest that the fault is a barrier to regional fluid flow. Our results also indicate that the permeability of the fault core is sufficiently low to result in effectively undrained behavior during slip, thus allowing dynamic processes including thermal pressurization and dilatancy hardening to affect slip behavior.
5.1. INTRODUCTION

Fluid flow and pressure in fault zones are two of the most important controls on fault strength and plate boundary motion. For the San Andreas Fault (SAF), both mechanical [Zoback et al., 1987; Hickman, 1991; Hickman and Zoback, 2004; Townend and Zoback, 2004] and thermal studies [Lachenbruch and Sass, 1980; Lachenbruch and McGarr, 1990; Fulton et al., 2004] imply that the fault is weak such that total resolved shear stress is on the order of earthquake stress drop (~10-20MPa or less) not the 50-100 MPa predicted by Byerlee’s Law and laboratory measurements.

High pore pressure reducing the effective stress has been invoked to explain the apparent weakness of many large fault zones [e.g. Hubbert and Rubey, 1959]. A common explanation for the apparent weakness of the SAF is that elevated fluid pressure is localized within the fault zone [Rice, 1992; Byerlee, 1990]. Fluid sources [e.g. Irwin and Barnes, 1975; Rice, 1992], porosity loss [e.g. Sleep and Blanpied, 1992], and dynamic weakening [e.g. Segall and Rice, 1995; Andrews, 2002] have all been presented as hypotheses for pore pressure generation. Evaluating each of these models requires knowledge of the permeability of the fault and surrounding material.

Models of the SAF have been constructed with large regional features of fluid and host rock properties and evaluated by sensitivity analysis [e.g. Saffer et al., 2003; Fulton et al., 2004; Fulton and Saffer, 2009]. These studies show that pore pressure development is highly dependent on both the permeability of the fault zone and difference of the permeability of the fault zone to the surrounding crust [e.g. Rice,
1992; Byerlee, 1990; Segall and Rice, 1995; Fulton and Saffer, 2009]. To better constrain modeling results and our understanding of the potential for pore pressure generation to explain fault weakness it is necessary to characterize the permeability of rocks within the crust and the fault zone.

A fault zone can act as either a fluid conduit or barrier [Caine et al., 1996]. Fault zone permeability has been established from surface samples on the Carbonaras fault, Spain [Faulkner and Rutter, 2003], Median Tectonic Line, Japan [Wibberly and Shimamoto, 2003], on thin fault zones from core samples on the Nojima fault, Japan [Lockner et al., 2000], Aegion Fault gouge [Sulem et al., 2004], or from subduction zone drilling. Wiersburg and Erzinger [2007] used $^3\text{He}/^4\text{He}$ during San Andreas Fault Observatory at Depth (SAFOD) drilling to conclude that the SAF was a barrier to flow. Measurements during drilling showed no over pressure in the fault zone [Zoback et al. 2010]. We report on laboratory experiments collected on in situ samples of fault gouge from the SAF to quantify the observations that the SAF is low permeability hydrologic barrier.

Gouge samples from SAFOD are unique among fault zone samples in that they are collected from depth and at in situ conditions on a thick large displacement fault zone. Our samples are chosen from two sections of fault gouge from the SAFOD core in the Central Deformation Zone (CDZ) a 2.6 m thick strand of the SAF occurring ~2.7 km below the surface. These sections are characterized by significant casing deformation during the SAFOD project. See Zoback et al. [2010] and references therein for a description of drilling and the samples.
5.2. METHODS

5.2.1 Experimental Apparatus

Experiments are conducted in either a 10,000 psi Temco triaxial core holder or 50 kN GDS load frame with a uniaxial pressure cell (Figure 5.1.) Uniaxial conditions are imposed as a zero radial strain boundary condition by surrounding the sample with a rigid steel ring (Figure 5.1a). A vertical load, \( \sigma_v \), is applied by a steel piston at up to 50 kN. Fluid pressure is delivered to the top and bottom of the sample by GDS pumps (Figure 5.1a). The fluid pressure is monitored at the base of the sample during constant rate of strain tests, and in each pump during flow-through tests.

In our triaxial experiments, axial stress, \( \sigma_a \), is kept equal to confining pressure, \( P_c \), in an isostatic condition such that \( \sigma_1 = \sigma_2 = \sigma_3 \). The pressure is applied via an ISCO 10,000 psi pump. The sample is jacketed with a viton rubber tube that fits over two end caps with a 1” diameter. Pore pressure is supplied from two GDS, 3 MPa pumps. They can be operated in tandem for flow through tests or with one pump removed in transient permeability tests. The entire assembly is housed in a temperature control box with control of \( \pm 0.1^\circ C \) at 30°C. The core holder can accommodate samples from 0.75” to 3” long and 1” in diameter, corresponding to \( \sigma_1 = \sigma_2 = \sigma_3 \approx 69 \) MPa.

5.2.2 Permeability Methods
Permeability experiments were conducted as transient pulse-decays, constant rate of strain (CRS) and flow-through tests (as both constant flow rate and constant pressure differential). For Interlab Comparison samples, we used deaerated, de-ionized water, and on fault zone samples we use deaerated water with fluid chemistry equivalent to the SAFOD borehole. All tests were conducted on 1” diameter samples with lengths of 25-40 mm and 15-20 mm for triaxial and uniaxial stress experiments, respectively.

Pulse-decay tests were preformed in the Temco triaxial core-holder at $P_c'$, of 5-60 MPa. Pore pressure was increased along with the isostatic stress to an initial value of 1 MPa. Pulse sizes ranged from 0.2 to 1 MPa from either an infinite upstream or small ($\sim 1.5 \text{ cm}^3$) reservoir into a downstream reservoir $\sim 1 \text{ cm}^3$. The pressure was monitored both upstream and downstream. Permeability and specific storage were calculated via curve matching [Hsieh et al., 1981]. Models of permeability and storage are compared with the response of the downstream reservoir with the curve that minimizes the difference between the data and the model (Figure 5.2).

Constant rate of strain and constant head permeability experiments were conducted in the GDS uniaxial load frame. The sample is deformed at a constant rate of displacement with the increase in pore pressure monitored [e.g. ASTM International, 2006; Saffer and McKiernan, 2005; Skarbek and Saffer, 2009]. The sample was allowed to back pressure for over 24 h before beginning the CRS test. An example of one experiment is given in Figure 5.3. As strain increases, the sample length decreases and an excess pressure builds above the background value. From the
strain rate, length, and excess pressure, permeability, $k$, can be calculated continuously by:

$$k = \frac{\nu \cdot \varepsilon \cdot l \cdot l_0}{2U_b}$$  \hspace{1cm} (1)

where $\nu$ is dynamic viscosity of water at 20°C, $\varepsilon$ is strain rate, $l$ is sample length, $l_0$ is initial sample length and $U_b$ is the excess pore pressure from the ambient state.

We calculate the volumetric compressibility, $m_v$, from the ratio of change in strain to effective stress during each increment. The coefficient of consolidation, $c_v$, can then be found from the relation:

$$c_v = km_v$$  \hspace{1cm} (2)

Constant pressure differential flow-through experiments, were conducted to confirm measurements from the transient methods. Permeability was calculated using Darcy’s law:

$$k = \nu \frac{Q}{A} \frac{l}{dP}$$  \hspace{1cm} (3)

where $Q$ is the flow rate and $dP$ is pressure differential across a sample of cross sectional area $A$. In all cases, multiple pressure differentials were used to minimize error in our calculation (Figure 5.4).

5.3. INTERLAB COMPARISON

To evaluate the reliability of our measurements and provide a set of laboratory standards, we conducted a series of experiments on samples of various permeabilities with multiple methods. These samples represent a coordinated effort by labs around
the world to develop a set of baseline comparisons so that measurements between labs can be rationalized [e.g. Lockner et al., 2009; http://www.geosc.psu.edu/~cjm/safod].

We conducted experiments on samples of Berea Sandstone, Crab Orchard Sandstone (commonly referred to as ‘Tennessee Sandstone’ in rock mechanics literature), and Wilkeson Sandstone. Interlab comparison samples were tested in the Temco apparatus under isostatic stress conditions of 5 MPa 10 MPa, 30 MPa, and 60 MPa, with multiple measurements at 30MPa and 60 MPa to evaluate any permeability hysteresis. Berea and Crab Orchard sandstones were cored in three orthogonal directions (T-B, N-S, E-W) while Wilkeson Sandstone was tested in one direction. See Table 5.1 for sample details and measured values. In both the Berea and Crab Orchard sandstones, the T-B sample represents across bedding flow.

The permeability of Berea Sandstone ranges from $4 \times 10^{-14} \text{ m}^2$ to $2.7 \times 10^{-15} \text{ m}^2$ from constant flow rate tests (Figure 5.5). Both the N-S and T-B samples show a decrease in $k$ from $P_c'$ of 5 MPa to 30 MPa. All three samples show considerable hysteresis when $P_c'$ is cycled between 30 and 60 MPa. In our tests, Berea Sandstone is anisotropic with the N-S sample decreasing in $k$ much less than the other two samples and is approximately one order of magnitude more permeable at 60 MPa $P_c'$. Published permeabilities for Berea sandstone are similar to our measured values. David et al. [1994] report a $k \sim 8 \times 10^{-14} \text{ m}^2$ at $P_c'=50 \text{ MPa}$.

Constant pressure differential experiments were run in each of the three orthogonal directions of the Crab Orchard Sandstone and the pulse-decay method was used on the N-S orientation. Permeability ranges from $2 \times 10^{-18} \text{ m}^2$ to $5.2 \times 10^{-20} \text{ m}^2$. 

122
Similar to Berea Sandstone, Crab Orchard decrease in permeability with increased $Pc'$ (Figure 5.6). The N-S and E-W sample decrease by about an order of magnitude with the T-B sample displaying more stress dependence. The high stress dependence is likely the result of the bedding in the sample and closing of permeable pathways. Crab Orchard Sandstone shows anisotropy and hysteresis in all of the flow-through tests. Pulse-decay tests yield a smaller permeability than flow-through at low $Pc'$ (Figure 5.6). When $Pc'$ reaches 60 MPa the results from both tests are similar. Keaney et al. [2004] report $k = 3 \times 10^{-18}$ m$^2$ at 20 MPa $Pc'$ from pulse tests while Benson et al. [2005] report a highly anisotropic and pressure dependent permeability for the Crab Orchard Sandstone. They find $k \sim 1 \times 10^{-19}$ m$^2$ perpendicular to bedding and $3 \times 10^{-19}$ m$^2$ with bedding. Benson et al. [2005] report a permeability anisotropy of $\sim 100\%$ from $Pc' = 5$ to 90 MPa, similar to our results.

Tests on the Wilkeson Sandstone were completed in the E-W orientation and as with the other two sandstones, considerable stress dependence is observed in $k$ with some hysteresis. Permeability starts at $3.3 \times 10^{-18}$ m$^2$ at $Pc' = 5$ MPa and evolves to $2 \times 10^{-19}$ m$^2$ at a $Pc' = 60$ MPa. Cycling of $Pc'$ shows considerable hysteresis similar to the other sandstones.

5.4. GEOLOGIC SETTING AND SAMPLE DESCRIPTION

The SAFOD project was located in central California in the creeping section of the SAF near Parkfield. The SAF is a $\sim 1300$ km long right lateral strike-slip fault that is seismogenic in the southern and northern segments and aseismic in the central portion. In the Parkfield region, the upper portion of the fault is in contact with
Tertiary sediments. At depth, the Pacific side is in contact with arkosic sandstone and conglomerates while on the continental side, the Great Valley Formation is adjacent to the fault with the Franciscian Complex further to the east (Figure 5.8a).

The SAFOD borehole is ~3km from the surface trace of the fault with the borehole vertical until ~1.5 km and then deviates to intersect the fault perpendicular to dip (Figure 5.8). A 200m thick damage zone containing fractured rock with low P and S wave velocity surrounds the fault zone [Zoback et al., 2010]. Drilling intersected two gouge zones, termed the Southwestern (SDZ) and Central Deformation Zones (CDZ) at ~2.7 km below the surface. Both of these zones underwent active casing deformation between October 2005 and June 2007, the time period between Phase 2 and 3 drilling. Caliper logs indicate ~90% of the total deformation occurred in the CDZ.

We conduct experiments on the 2.6 m thick CDZ. Our samples are from two sections (4, 5) of the side lateral Hole G, Run 4 of Phase 3 drilling. The gouge is foliated with a wavy-fabric, highly altered and sheared throughout with surface striations (Figure 5.8b). The matrix is composed of particles <10 μm is diameter with ~5% porphyroclasts up to several cm in diameter. Porphyroclast lithogies are serpentinite, very fine-grained sandstone, siltstone and white vein fragments.

5.5. PERMEABILITY OF THE SAN ANDREAS FAULT

Flow-through and CRS tests indicate that the permeability of the CDZ is low. CRS experiments on core intervals Hole G, Run 4, section 4 and 5 and a repeat of G, 4, 4 show remarkable similarity in $k$. For all samples the permeability starts at $\sim 10^{-17}$
m² at $Pc' \sim 1$ MPa and decreases rapidly to $\sim 10^{-19}$ m² by $\sigma_v' = 20$ MPa. Below 20 MPa, $k$ shows little to no change with increased $\sigma_v'$ for experiments U163 and U164 and decreases slightly in U173 (Figure 5.9). The values at low $Pc'$ (<5 MPa) are likely the result of fluid pressurization during the early stages of the experiment. To check the viability of CRS tests our values are compared to flow-through tests at low stress. At 5 MPa $Pc'$, $k = 2.2 \times 10^{-19}$ m² an equivalent value to the CRS test. At 10 MPa permeability is slightly lower, but similar to CRS (Figure 5.9). See Table 5.2 for a complete list of all experiments and methods for tests on the CDZ.

The specific storage of the gouge can be calculated from

$$Ss = g \rho_f (m_v + \phi \beta)$$  \hspace{1cm} (4)

where $g$ is the gravitational constant, $\rho_f$ fluid density, $\phi$ porosity and $\beta$ is the fluid compressibility. As with $k$ and compressibility, $Ss$ rapidly decreases until $\sigma_v' \sim 20$ MPa then remains near constant at $Ss \approx 10^{-5}$ m⁻¹ (Figure 5.10).

The initial low values of storage and permeability are the result of stiffening of the sample at the onset of the experiment. At low stress and strain, Young’s Modulus, $E$, changes considerably with strain (Figure 5.11). After $\sigma_v' > 20$ MPa, $E$ remains near constant at $\sim 1.5$ GPa⁻¹, a typical value for sediment and mud samples. The amount of strain required to reach $\sigma_v' = 20$ MPa varies from test to test (Figure 5.11). Based on this stiffening, only values for $k$ and $Ss$ after a constant $E$ has been achieved should be used.

### 5.5.1 Comparison with other data
Our data agree with other fault studies from both surface sample and core measurements. Helium isotope studies indicate that the SAF is a barrier to regional fluid flow [Wiersberg and Erzinger, 2007]. Work on surface samples of the SAF in Cienga Valley found that permeability ranged from $\sim 10^{-20}$ to $10^{-21}$ from $\sim 10$-20 MPa until 100 MPa and decreased to $\sim 10^{-22}$ m$^2$ by 200 MPa [Morrow et al., 1981]. Morrow et al. [1984] tested a variety of gouges from different portions of the SAF with varied mineralogy ranging from serpentinite to montmorillonite-rich. They found that permeability ranged from $10^{-18}$ m$^2$ to $10^{-21}$ m$^2$. Surface samples from other fault zones show similar values [e.g. Faulkner and Rutter, 2003; Sulem et al., 2004]. Wibberley and Shimamoto [2003] show that the permeability and structure of the Median Tectonic Lines is complex both in and around the fault. At it’s core the Median Tectonic Line has $k \sim 10^{-19}$ m$^2$ increasing as much as 4 orders of magnitude away from the fault. Experiments by Faulkner and Rutter [1998] indicate that fault zone permeability can be highly dependent on foliation. Permeability in the direction of foliation is 2-3 orders of magnitude larger in phyllosilicate-rich samples than permeability across foliation. Our experiments are conducted perpendicular to the fault zone and the wavy foliation of the gouge (Figure 5.8b).

5.5.2 Implications

The permeability of fault zones is key in constraining models of earthquake rupture and slow slip. Thermal pressurization has been hypothesized as a mechanism for rupture [i.e. Andrews, 2002]. Segall and Rice [2006] show for a low permeability fault, that thermal pressurization alone is insufficient rupture propagation. Andrews...
[2002] assumes a more permeable bulk fault zone with \( k = 5 \times 10^{17} \text{ m}^2 \). From these values it is concluded that frictional heating can raise the fluid pressure and reduce friction.

Dilantancy hardening has been invoked to explain the regulation of glacial slip [Moore and Iverson, 2002], aseismic creep and slow-slip [e.g. Seront et al., 1998; Rubin, 2008; Samuelson et al., 2009; Segall et al., in press]. The coupled low permeability and large thickness of the SAF the fault zone can cause the fault to behave essentially undrained. Permeability in the range of \( 10^{-20} \) to \( 10^{-21} \text{ m}^2 \) coupled with dilation could depressurize the fault by as much as 50% [Samuelson et al., 2009]. Our data show that the SAF can act undrained which allows for the possibility of thermal weakening or dilatancy hardening. A fault zone on the order of \( k \sim 10^{-20} \text{ m}^2 \) surrounded by fractured crust would act as a barrier to fluid and affect the regional hydrology.

**5.6 CONCLUSIONS**

Permeability experiments in several configurations show that the San Andreas fault has a low permeability and would act as a regional barrier to fluid migration. The low permeability of the Central Deformation Zone of the fault would facilitate coupled hydro-mechanical processes such as thermal weakening and dilatancy hardening. We also present the permeability on a suite of known samples for direct comparison of our results with other studies. We find that the permeability of the main strand of the SAF has a permeability of \( \sim 10^{-20} \text{ m}^2 \).
5.7 ACKNOWLEDGEMENTS

We thank Steve Swavley for technical assistance and Sam Haines for SEM images.

REFERENCES


ASTM International (2006), Standard test method for one-dimensional consolidation properties of saturated cohesive soils using controlled-strain loading (Standard D4186-06), in Annual Book of ASTM Standards: Soil and Rock (I), vol. 04.08, West Conshocken, PA.


<table>
<thead>
<tr>
<th>Test</th>
<th>Material</th>
<th>Orientation</th>
<th>Method</th>
<th>$P_c'$ (MPa)</th>
<th>$k$ (m$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>T 101</td>
<td>Berea</td>
<td>N-S</td>
<td>Const. Flow</td>
<td>5</td>
<td>3.66$x10^{-14}$</td>
</tr>
<tr>
<td>T 101</td>
<td>Berea</td>
<td>N-S</td>
<td>Const. Flow</td>
<td>10</td>
<td>3.54$x10^{-14}$</td>
</tr>
<tr>
<td>T 101</td>
<td>Berea</td>
<td>N-S</td>
<td>Const. Flow</td>
<td>30</td>
<td>3.69$x10^{-14}$</td>
</tr>
<tr>
<td>T 101</td>
<td>Berea</td>
<td>N-S</td>
<td>Const. Flow</td>
<td>60</td>
<td>3.54$x10^{-14}$</td>
</tr>
<tr>
<td>T 101</td>
<td>Berea</td>
<td>N-S</td>
<td>Const. Flow</td>
<td>30</td>
<td>3.45$x10^{-14}$</td>
</tr>
<tr>
<td>T 101</td>
<td>Berea</td>
<td>N-S</td>
<td>Const. Flow</td>
<td>60</td>
<td>2.6$x10^{-14}$</td>
</tr>
<tr>
<td>T 101</td>
<td>Berea</td>
<td>N-S</td>
<td>Const. Flow</td>
<td>30</td>
<td>2.62$x10^{-14}$</td>
</tr>
<tr>
<td>T 101</td>
<td>Berea</td>
<td>N-S</td>
<td>Const. Flow</td>
<td>60</td>
<td>2.73$x10^{-14}$</td>
</tr>
<tr>
<td>T 98</td>
<td>Berea</td>
<td>E-W</td>
<td>Const. Flow</td>
<td>5</td>
<td>4.35$x10^{-14}$</td>
</tr>
<tr>
<td>T 98</td>
<td>Berea</td>
<td>E-W</td>
<td>Const. Flow</td>
<td>10</td>
<td>2.96$x10^{-14}$</td>
</tr>
<tr>
<td>T 98</td>
<td>Berea</td>
<td>E-W</td>
<td>Const. Flow</td>
<td>30</td>
<td>2.46$x10^{-14}$</td>
</tr>
<tr>
<td>T 98</td>
<td>Berea</td>
<td>E-W</td>
<td>Const. Flow</td>
<td>60</td>
<td>5.06$x10^{-15}$</td>
</tr>
<tr>
<td>T 98</td>
<td>Berea</td>
<td>E-W</td>
<td>Const. Flow</td>
<td>30</td>
<td>6.62$x10^{-15}$</td>
</tr>
<tr>
<td>T 98</td>
<td>Berea</td>
<td>E-W</td>
<td>Const. Flow</td>
<td>60</td>
<td>4.57$x10^{-15}$</td>
</tr>
<tr>
<td>T 98</td>
<td>Berea</td>
<td>E-W</td>
<td>Const. Flow</td>
<td>30</td>
<td>5.83$x10^{-15}$</td>
</tr>
<tr>
<td>T 98</td>
<td>Berea</td>
<td>E-W</td>
<td>Const. Flow</td>
<td>60</td>
<td>4.51$x10^{-15}$</td>
</tr>
<tr>
<td>T 99</td>
<td>Berea</td>
<td>T-B</td>
<td>Const. Flow</td>
<td>5</td>
<td>2.19$x10^{-14}$</td>
</tr>
<tr>
<td>T 99</td>
<td>Berea</td>
<td>T-B</td>
<td>Const. Flow</td>
<td>10</td>
<td>1.24$x10^{-14}$</td>
</tr>
<tr>
<td>T 99</td>
<td>Berea</td>
<td>T-B</td>
<td>Const. Flow</td>
<td>30</td>
<td>6.82$x10^{-15}$</td>
</tr>
<tr>
<td>T 99</td>
<td>Berea</td>
<td>T-B</td>
<td>Const. Flow</td>
<td>60</td>
<td>5.23$x10^{-15}$</td>
</tr>
<tr>
<td>T 99</td>
<td>Berea</td>
<td>T-B</td>
<td>Const. Flow</td>
<td>30</td>
<td>3.95$x10^{-15}$</td>
</tr>
<tr>
<td>T 99</td>
<td>Berea</td>
<td>T-B</td>
<td>Const. Flow</td>
<td>60</td>
<td>3.38$x10^{-15}$</td>
</tr>
<tr>
<td>T 99</td>
<td>Berea</td>
<td>T-B</td>
<td>Const. Flow</td>
<td>30</td>
<td>3.13$x10^{-15}$</td>
</tr>
<tr>
<td>T 99</td>
<td>Berea</td>
<td>T-B</td>
<td>Const. Flow</td>
<td>60</td>
<td>2.67$x10^{-15}$</td>
</tr>
<tr>
<td>T 121</td>
<td>Crab Orchard</td>
<td>N-S</td>
<td>Pulse</td>
<td>10</td>
<td>1.0$x10^{-19}$</td>
</tr>
<tr>
<td>T 121</td>
<td>Crab Orchard</td>
<td>N-S</td>
<td>Pulse</td>
<td>30</td>
<td>1.85$x10^{-19}$</td>
</tr>
<tr>
<td>T 121</td>
<td>Crab Orchard</td>
<td>N-S</td>
<td>Pulse</td>
<td>60</td>
<td>1.1$x10^{-19}$</td>
</tr>
<tr>
<td>T 121</td>
<td>Crab Orchard</td>
<td>N-S</td>
<td>Pulse</td>
<td>30</td>
<td>2.0$x10^{-19}$</td>
</tr>
<tr>
<td>T 121</td>
<td>Crab Orchard</td>
<td>N-S</td>
<td>Pulse</td>
<td>60</td>
<td>1.2$x10^{-19}$</td>
</tr>
<tr>
<td>T 108</td>
<td>Crab Orchard</td>
<td>N-S</td>
<td>Const. Head</td>
<td>5</td>
<td>2.09$x10^{-18}$</td>
</tr>
<tr>
<td>T 108</td>
<td>Crab Orchard</td>
<td>N-S</td>
<td>Const. Head</td>
<td>10</td>
<td>1.18$x10^{-18}$</td>
</tr>
<tr>
<td>T 108</td>
<td>Crab Orchard</td>
<td>N-S</td>
<td>Const. Head</td>
<td>30</td>
<td>6.21$x10^{-19}$</td>
</tr>
<tr>
<td>T 108</td>
<td>Crab Orchard</td>
<td>N-S</td>
<td>Const. Head</td>
<td>60</td>
<td>1.98$x10^{-19}$</td>
</tr>
<tr>
<td>T 108</td>
<td>Crab Orchard</td>
<td>N-S</td>
<td>Const. Head</td>
<td>30</td>
<td>3.02$x10^{-19}$</td>
</tr>
<tr>
<td>T 108</td>
<td>Crab Orchard</td>
<td>N-S</td>
<td>Const. Head</td>
<td>60</td>
<td>2.0$x10^{-19}$</td>
</tr>
<tr>
<td>T 108</td>
<td>Crab Orchard</td>
<td>N-S</td>
<td>Const. Head</td>
<td>30</td>
<td>3.55$x10^{-19}$</td>
</tr>
<tr>
<td>T 108</td>
<td>Crab Orchard</td>
<td>N-S</td>
<td>Const. Head</td>
<td>60</td>
<td>1.73$x10^{-19}$</td>
</tr>
<tr>
<td>T 109</td>
<td>Crab Orchard</td>
<td>T-B</td>
<td>Const. Head</td>
<td>5</td>
<td>1.23$x10^{-18}$</td>
</tr>
<tr>
<td>T 109</td>
<td>Crab Orchard</td>
<td>T-B</td>
<td>Const. Head</td>
<td>10</td>
<td>1.96$x10^{-19}$</td>
</tr>
<tr>
<td>T 109</td>
<td>Crab Orchard</td>
<td>T-B</td>
<td>Const. Head</td>
<td>30</td>
<td>9.39$x10^{-20}$</td>
</tr>
<tr>
<td>T 109</td>
<td>Crab Orchard</td>
<td>T-B</td>
<td>Const. Head</td>
<td>60</td>
<td>5.26$x10^{-20}$</td>
</tr>
<tr>
<td>T 109</td>
<td>Crab Orchard</td>
<td>T-B</td>
<td>Const. Head</td>
<td>30</td>
<td>8.0$x10^{-20}$</td>
</tr>
<tr>
<td>Test</td>
<td>Sample</td>
<td>Method</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>-------</td>
<td>--------</td>
<td>---------------------</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>T162</td>
<td>G 4, 5</td>
<td>Triax, Const. Head</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>U163</td>
<td>G 4, 5</td>
<td>Uniax, CRS</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>U164</td>
<td>G 4, 4</td>
<td>Uniax, CRS</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>U173</td>
<td>G 4, 4</td>
<td>Uniax, CRS</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 5.1. Interlab comparison results.

Table 5.2. Experiment list.
Figure 5.1. Permeability apparatuses. (a) Uniaxial pressured conditions are maintained by placing the sample in a solid steel ring inside of a pressure vessel. Vertical stress is applied via a steel ram with a flow distribution cap. Pore pressure is controlled independently at the top at the bottom of the sample. (b) Triaxial core holder. Axial stress and confining pressure are kept equal to maintain isostatic conditions. Pore pressure is delivered through end caps at each end with a porous metal frit attached.
Figure 5.2. Pulse decay curve matching. (a) Curves are matched for three reservoir sizes when possible. The curve for with the smallest error for storage and permeability is chosen. In cases when the reservoir storage is ≥ sample storage the specific storage of the sample is unconstrained.
Figure 5.3. Constant rate of strain uniaxial permeability test. A constant displacement rate is imposed on the top of a sample, which is drained to a pump holding a constant pressure. The pore pressure build-up from an initial value (0.5 MPa in this test) is monitored at the bottom of the sample. (inset) Zoomed version of the pore pressure increase.
Figure 5.4. Permeability calculation from constant flow test. Pressure differential, $dP$, across the sample is controlled and the flow rate, $Q$ is monitored in two pumps. The slope of the best-fit line is proportional to the permeability, $k$. In all cases multiple $dP$ are used to minimize error with zero flow assumed at $dP = 0$. 
Figure 5.5. Permeability of Berea Sandstone in three orthogonal directions. Permeability is calculated from holding the flow rate constant at an effective pressure of 5-60 MPa. Pressure is cycled between 30 and 60 MPa to monitor the hysteresis of the sample. Dashed lines indicate the order of the measurement. The T-B sample is across formation bedding.
**Figure 5.6.** Permeability of Crab Orchard (often called Tennessee) Sandstone in three orthogonal directions. Permeability is calculated from holding the pressure differential constant at an effective pressure of 5-60 MPa and in a repeat experiment, from the pulse decay method. Pressure is cycled between 30 and 60 MPa to monitor the hysteresis of the sample. Dashed lines indicate the order of the measurement. The T-B sample is across formation bedding.
Figure 5.7. Permeability of the Wilkeson Sandstone in one direction. Permeability is calculated from holding the pressure differential constant at an effective pressure of 5-60 MPa. Pressure is cycled between 30 and 60 MPa to monitor the hysteresis of the sample. Dashed lines indicate the order of the measurement. No bedding is present in our sample.
Figure 5.8. (a) Geologic cross-section of the SAFOD project, from Zoback et al. [2010]. The borehole deviates to intersect the zones at ~90°. Fault zones are marked in red at ~2.7 km depth. Earthquakes are indicated by circles. The first fault zone is the SDZ and the second is the CDZ. (b), (c) SEM images of the CDZ from cuttings of permeability sample preparation.
Figure 5.9. Permeability, $k$, drops rapidly in the three CRS tests until effective stress ~20 MPa then remains near constant. Triaxial flow-through tests show equal permeability at 5 MPa and lower values at 10 MPa.
Figure 5.10. Specific storage, $S_s$, with vertical stress, $\sigma'_v$. $S_s$ rapidly decreases at low stress and then decreases slightly after $\sigma'_v \sim 20$ MPa. Both experiments show consistent values at $10^{-5}$ m$^{-1}$.
Figure 5.11. Young’s Modulus, $E$, rapidly increases until 20 MPa and then remains constant. Fiducial lines are given for reference values of $E$. The amount of strain varies from test to test to reach constant $E$. $1.5 \text{ GPa}^{-1}$ corresponds is the best-fit line of experiment U164 at $\sigma_v' > 40 \text{ MPa}$. 
APPENDIX A

This appendix represents work in which I am a co-author but not the primary author. This paper was published in *Meso-Scale Shear Physics in Earthquake and Landslide Mechanics* and deals with localization in fault zones and relates to Chapters 2-4 of my thesis.


Strain Localization in Granular Fault Zones at Laboratory and Tectonic Scales

C. Marone & A. P. Rathbun

*Dept. of Geosciences and Center for Geomechanics, Geofluids, and Geohazards, The Pennsylvania State University, University Park PA, USA*
ABSTRACT: We present results from laboratory experiments and a numerical model for frictional weakening and shear localization. Experiments document strain localization in sheared layers at normal stresses of 0.5 to 5 MPa, layer thicknesses of 3 to 10 mm, and imposed slip velocities of 10 to 100 μm/s. Passive strain markers and the response to load perturbations indicate that the degree of shear localization increases for shear strains $\gamma$ of $0.15 < \gamma < 1$. Our numerical model employs rate-state friction and uses 1D elasto-frictional coupling with radiation damping. We interrogate the model frictional behavior by imposing perturbations in shear velocity at the fault zone boundary. The spatial distribution of shear strain depends strongly on frictional behavior of surfaces within the shear zone. We discuss the onset of strain localization and the width of active shear strain for conditions relevant to earthquake faulting and landslides.

1 INTRODUCTION

Laboratory and field evidence indicate that strain localization is accompanied by significant changes in hydraulic and mechanical properties of rocks (e.g., Wood 2002, Song et al. 2004, Rice 2006). Strain localization occurs at a broad range of scales and involves both formation of faults and, upon continued shear, confinement of shear to narrow bands within the wear and gouge materials that constitute the fault zone. Of particular interest is the connection between strain localization and the transition from stable to unstable frictional sliding within shear zones of finite width (e.g., Anand & Gu 2000, Rice & Cocco 2007).

2 SHEAR LOCALIZATION IN GRANULAR LAYERS

In this paper we focus on layers composed of granulated rock. Granular layers were sheared in a biaxial deformation apparatus using the double-direct shear configuration. Details of the testing apparatus and experimental procedures are reported in Rathbun et al. (2008). Layers were initially 3 to 10 mm-thick and we imposed slip velocities of 10 to 100 μm/s at the layer boundary. Normal stress was held constant during shear via a fast-acting servo-hydraulic control mechanism. We discuss experiments conducted at normal stresses in the range 0.5 to 5 MPa, which is high enough to result in inelastic yield at grain to grain contacts, on the upper end, and low enough to inhibit grain crushing, on the lower end.

2.1 Dilation as a proxy for shear localization

Previous studies of granular layers have established that upon shear loading, shear stress rises linearly before undergoing a progressive transition from elastic to inelastic behavior (e.g., Anthony & Marone 2005). Inelastic yield is associated with grain rearrangement, compaction, and bulk shear strain of the layer (e.g., Marone 1998). Figure 1 illustrates this behavior for a granular layer that was initially 10 mm thick and sheared at a normal stress of 1 MPa. Note the steep rise in shear stress followed by strain hardening and a transition to steady frictional sliding at a shear strain of $\sim 0.5$. Layers compact during the initial rise in shear stress and dilation beings at a normalized stress (friction) level of 0.35 to 0.4 (Fig. 1). In this experiment we evaluated the effects of changes in slip velocity at the layer boundary. The systematic variation in friction seen throughout the experiment is associated with step changes in loading rate between 10 and 30 μm/s, as discussed further below.
Figure 1. Complete stress-strain curve for a granular layer sheared at a 1 MPa normal stress. Initial layer thickness was 10 mm. Normalized shear stress is plotted versus engineering shear strain computed from incremental displacement at the layer boundary divided by instantaneous layer thickness. Note the steep, quasi-linear, initial rise in shear stress followed by hardening and fully-mobilized shear at strains of 0.4 to 0.5. Beginning at a shear strain of 0.3, the shearing rate at the layer boundary was toggled between 10 and 30 μm/s. Inset shows detail of the friction response to step changes in slip velocity. The material is glacial till, (Caesar till, Ohio, USA) which is a mixed-size granular material (Particle size D: Dmax = 1 mm, D50 ~ 0.3 mm, 90% of the particles > 0.1 mm) similar to natural fault gouge (after Rathbun et al. 2008).

In granular layers, the transition from initial strain hardening to steady-state (fully mobilized) frictional sliding is associated with the development of localized shear (e.g., Logan et al. 1979, Marone et al. 1992). Recently, Rathbun et al. (2008) showed that stress perturbations, and layer dilation, provide a more precise measure of the degree of shear localization. They showed that small perturbations in the shear stress level during creep friction tests provide a proxy for shear localization (Fig. 2). Creep friction tests were carried out at a constant shear stress level below that for stable
sliding. Using this approach, the degree of shear localization can be assessed with the parameter $\Delta h^*$, which is the layer dilation for a unit increase in shear stress. We measured $\Delta h^*$ as a function of shear strain for a range of conditions (Fig. 2). Dilation scales with layer thickness below a critical value of shear strain. For shear strains $\lesssim 0.5$, thicker layers exhibit larger values of $\Delta h^*$ than thinner layers (Fig. 2). However for shear strains greater than $\sim 1$, dilation is independent of layer thickness. These data show that, initially, shear is distributed across the full thickness of the layer, but that shear becomes localized beyond a critical shear strain. The data of Figure 2 indicate that shear is fully localized by shear strains of roughly unity.

Figure 2 shows additional details of the relationship between shear stress, shear strain, and layer thickness. This figure shows stress-strain curves for representative experiments and one data set for changes in layer thickness as a function of shear strain (experiment p1025). Layer dilation occurs early in the strain history and then the layers compact slightly before reaching a steady level, consistent with a critical state, for shear strains of 0.3 and greater (Fig. 3).
2.2 Rate/State friction and shear localization

Slip velocity step tests have emerged as a powerful tool for interrogating friction constitutive behavior (Dieterich 1979, Ruina 1983, Scholz, 1998). A large body of literature shows that frictional strength of a wide range of materials exhibits two responses to a step increase in the imposed loading rate (e.g., Dieterich & Kilgore 1994, Tullis 1996, Marone, 1998). First, there is an instantaneous change in frictional resistance of the same sign as the velocity change. This is referred to as the friction direct effect and it is described by the friction parameter $a$. Figure 4 defines the key parameters and outlines the rate and state friction equations. The direct effect is followed by a gradual evolution of strength, scaled by the friction parameter $b$ (Fig. 4). The evolution effect is typically of the same as the change in velocity (Fig. 1). Existing studies show that the evolution effect occurs over a characteristic slip distance, $D_e$ (sometimes referred to as L), for initially-bare solid surfaces or a characteristic strain for layers of granular/clay particles (e.g., Marone 1998). The values of $D_e$ are typically larger for shear within a granular layer than for shear between solid surfaces (e.g., Marone & Kilgore 1993, Marone et al. 2009).

![Figure 3](image)

Figure 3. Complete stress strain curves for five experiments along with a representative data set (experiment p1025) for changes in layer thickness as a function of shear strain. The layer thickness started at 10 mm in experiment p1025 and the thickness was measured continuously during shear with a DCDT. Note that dilation occurs during the initial increase in stress but that compaction begins prior to fully-mobilized shear within the granular layer.
151

Figure 4. The equations describing the rate and state friction constitutive law along with a schematic showing this behavior.

To the extent that the friction evolution effect is truly driven by shear displacement, step velocity increases and decreases are expected to yield symmetric behavior. Such symmetry was found by Ruina (1983) and Marone et al. (1990). Other studies have favored a model in which friction evolution occurs over a characteristic time (Beeler et al. 1994, Sleep 1997). Finally, a large number of works have evaluated only velocity increases or decreases, without considering the issue of symmetry (e.g., Marone & Kilgore 1993). Few studies have systematically evaluated symmetry of the friction response to changes in loading velocity.

2.3 Mechanics of the critical slip distance for friction of granular materials

For solid surfaces in contact, the critical slip distance for friction evolution can be thought of in terms of the asperity contact lifetime, given by the contact size divided by the average slip rate (Rabinowicz 1951, Dieterich 1979). When coupled with the adhesive theory of friction (e.g., Bowden & Tabor 1950), in which asperity strength (and size) is proportional to time of contact (lifetime), this model predicts that frictional strength during steady-sliding should decrease with increasing slip velocity, because contact lifetime (hence strength) is inversely proportional to sliding velocity. In the context of a velocity step test, the critical friction distance is the slip necessary to replace contacts with a lifetime given by the initial velocity by contacts corresponding to the final velocity (Dieterich 1979, Ruina 1983).

For granular materials the situation is slightly more complex. The model for solid friction can be applied directly for asperity contacts between grains. However, granular interactions and stress transmission via particle contacts lead to a second characteristic length scale, in addition to the asperity contact junction size. The work by Marone & Kilgore (1993) shows that the critical slip distance for granular shear depends on both the particle size and the shear localization dimension (Fig. 5).
Figure 5. Measurements of the critical slip distance for granular layers as a function of shear strain. Data are shown for three particle size distributions. Note that the critical slip distance is greatest for larger particles at all values of shear strain. **Coarse:** Ottawa sand ASTM C-190 all particles are 600 to 800 μm. **Fine:** Silcosil 400 mesh (US Silica Co.) with median and maximum diameter of 1.4 and 10 μm, respectively. **Fractal:** given by \( N(n) = b n^{-D} \), where \( N(n) \) is the number of particles of size \( n \), \( b \) is a constant and \( D \) is the fractal dimension 2.6, made using particles in the range < 45 μm to 700 μm. Data from Marone & Kilgore (1993).

The data of Figure 5 show measurements of the critical slip distance as a function of shear strain for granular layers sheared in the double-direct shear geometry. The average particle sizes ranged from 700 μm (Coarse) to 5 μm (Fine); fractal is a power-law size distribution between 45 and 720 μm. The critical slip distance is greatest for larger particles at all values of shear strain (Fig. 5). Fine particles, with an average size that is roughly 100 times smaller than the coarse particles, have a critical slip distance that is roughly 10 times smaller than the coarse particles, consistent with the expected scaling between contact junction dimension and particle diameter. It is important to note, however, that the difference in \( D_c \) values for coarse and fine particles is small compared to the observed evolution of \( D_c \) with shear strain (Fig. 5). The \( D_c \) values for layers of both coarse and fine particles decrease by more than 100% of the final value at shear strains greater than ~ 7. The decrease in \( D_c \) with shear strain is consistent with the effects of shear localization.

### 2.4 Observations of shear localization

Laboratory investigations of shear localization often include post-experiment examination of preserved microstructures (e.g., Mair & Marone 1999). In the experiments described here, we used passive markers in some experiments to record the strain distribution across layers (Figure 6). The markers were constructed with blue sand grains. Following the shear experiment, layers were impregnated with
epoxy and then cut parallel to the shear direction. Figure 6 shows a layer that was sheared, top to the left, to a strain of 3.9. The layer had three markers that were initially vertical in the orientation of the photograph. The original image is shown below a copy (above) that has been marked to highlight the offset marker. Note that: 1) the marker is offset primarily along a zone near the center of the layer and 2) that the segments above and below the primary offset show distinct curvature. This curvature indicates a progressive localization process prior to development of the main shear zone in the center of the layer. By measuring offset of the top and bottom limbs of the marker, we calculated a shear strain of 3.25 along the shear zone at the center of the layer. Based on the total shear strain of 3.9, we estimate initiation of this shear band at a shear strain of 0.65, which is within the range indicated by our layer dilation measurements (Fig 2).

Figure 6. Thin section of a granular layer that was subject to a shear strain of 3.9. The layer contains three passive markers that were initially vertical, in this orientation, formed by darker particles. Top image is annotated to show shear of the central marker. Lower image is unmarked photograph.

Figure 7 is a schematic illustration of two modes by which shear could become localized. In panels a-c the marker is first subject to uniform strain and then cut by a shear zone. In this scenario, shear localizes abruptly. The markers are first rotated by simple shear and then cut and offset along a narrow zone at the center of the layer. Panels d-f show a more progressive localization process (Fig. 7). The markers are initially subject to uniform strain, but localization occurs gradually and on several surfaces near the center of the zone. The markers are bent into an arcuate shape by progressively greater strain concentration with increasing distance from the layer boundaries (Fig. 7). Eventually, the strained markers are offset along a primary shear band. This type of localization process, with a gradual transition from pervasive to localized strain is consistent with our dilation measurements, which indicate progressively greater localization over the range of macroscopic shear strain from 0.15 to 1.0. Moreover, thin sections from our experiments (Fig. 6) indicate a progressive localization process, like panels d-f of Figure 7, rather than an abrupt transition.
3 A NUMERICAL MODEL FOR FRICTIONAL WEAKENING AND SHEAR LOCALIZATION

To address shear localization in tectonic fault zones and to improve our understanding of the scaling problem associated with applying laboratory observations to faults in Earth's crust, we employ a numerical model. The model describes frictional shear in a fault zone composed of multiple, parallel surfaces that obey rate and state friction (Fig. 8). The model used here is based on that described by Marone et al. (2009). We extend that model and focus on coupling between friction properties and shear localization.

3.1 Elasto-Frictional Model for a Fault Zone of Finite Thickness

A typical tectonic fault zone consists of a highly damaged zone surrounded by progressively less damaged country rock (e.g., Chester & Chester 1998). Thus, in the context of the seismic cycle, the zone of slip deformation is defined by a critical fracture density, above which slip and deformation occurs, and below which the rock behaves as intact material. One could
Imagine a model in which this critical fracture density depended on strain rate and other factors, such that the effective fault zone width varied throughout the seismic cycle, but we make the simplifying assumption of constant width $T$ (Fig. 8).

Figure 8. Fault zone model and schematic of shear zone composed of multiple sub-parallel surfaces. The fault zone width is $T$. $K_{\text{ext}}$ represents elastic stiffness of the crust surrounding the fault. $K_{\text{int}}$ represents elastic coupling between surfaces in the model, which are separated by distance $h$. The model is symmetric about the center.

Within the model fault zone, shear may occur on one or more sub-parallel surfaces (Fig. 8). Our aim is to investigate spatio-temporal complexity of shear localization in a fault zone that experiences a rapid change in imposed slip rate; for example due to earthquake propagation into the region of interest. Therefore we employ a simplistic geometry and elastic model. As an initial condition, we assume homogeneous creep within the fault zone, such that all surfaces are slipping at a background rate. We investigate the fault response to perturbations in slip rate imposed at the fault zone boundary at $\pm T/2$ relative to the center of the zone (Fig. 8).

Potential slip surfaces within the fault zone interact via elasto-frictional coupling. Stress is transmitted between surfaces only when: 1) the frictional strength of a surface exceeds the current stress level, or 2) a surface slips and its strength changes. In the models described here, stresses and frictional strengths are initially equal on all surfaces. Slip surfaces obey laboratory-based rate and state friction laws, and we
focus here on the case of state evolution via the Ruina law (Dieterich 1979, Ruina 1983). A one-dimensional elastic model is used with radiation damping to solve the equations of motion.

Each surface $i$ in the model shear zone obeys rate and state frictional behavior, such that friction $\mu_i$ is a function of state $\theta_i$ and slip velocity $v_i$ according to:

$$\mu_i(\theta_i, v_i) = \mu_0 + a \ln \left( \frac{v_i}{v_0} \right) + b \ln \left( \frac{v_0 \theta_i}{L} \right)$$  

(1)

where $\mu_0$ is a reference friction value at slip velocity $v_0$, and the parameters $a$, $b$, and $L$ are empirically-derived friction constitutive parameters (e.g., Marone 1998). Note that we use $L$ for the model critical slip distance, rather than $D_c$, which is the effective parameter measured from laboratory experiments. Tectonic fault zones are likely to include spatial variations of the friction constitutive parameters within the shear zone, and thus we allow such behaviors.

The model includes $n_s$ parallel surfaces, where $i = 0$ is at the fault zone boundary. Surfaces are coupled elastically to their neighbors via stiffness $K_{int}$. We assume $K_{int} = G/h$, where $G$ is shear modulus and $h$ is layer spacing (Fig. 8) and use $G = 30$ GPa. We assume that remote tectonic loading of the shear zone boundary is compliant relative to $K_{int}$ and take $K_{int}/K_{ext}$ equal to 10.0. This is equivalent to assuming a constant spacing between surfaces and means that wider shear zones, with more internal surfaces, are effectively more compliant than narrower zones. Another approach would be to take $K_{int}/K_{ext}$ equal to the number of surfaces in the shear zone. Details of the parameters used are reported in Table 1.

Table 1. Model parameters. For all cases, $G = 30$ GPa, $\sigma = 100$ MPa, $K_{int} = G/h$; $K_{int}/K_{ext}$ =10; $v_0 = 1e^{-6}$ m. $n_s/2$ is the number of surfaces in the fault zone half width $T/2$.

<table>
<thead>
<tr>
<th>a</th>
<th>b</th>
<th>L (m)</th>
<th>h (m)</th>
<th>$K_{ext}/\sigma_n$ (m$^{-1}$)</th>
<th>$n_s/2$</th>
<th>T (m)</th>
<th>v (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.012</td>
<td>0.016</td>
<td>1e-5</td>
<td>6e-3</td>
<td>5e4</td>
<td>30</td>
<td>0.60</td>
<td>0.01</td>
</tr>
</tbody>
</table>

We analyze friction state evolution according to:

$$\frac{d\theta_i}{dt} = -\frac{v_i \theta_i}{L} \ln \left( \frac{v_i \theta_i}{L} \right). \text{ (Ruina Law)}$$  

(2)

Frictional slip on each surface satisfies the quasi-dynamic equation of motion with radiation damping (Rice 1993):

$$\mu_i = \frac{\tau_0}{\sigma_n} - \frac{G}{2\beta \sigma_n} (v_i - v_{pl}) + k(v_{pl} t - v_i)$$  

(3)

where $\mu_i$ is the frictional stress, $\tau_0$ is an initial stress, $\beta$ is shear wave speed, $\sigma_n$ is normal stress, $k$ is stiffness divided by normal stress, and $t$ is time. Differentiating Equations 1 and 3 with respect to time and solving for $dv_i/dt$ yields:
\[
\frac{dv_i}{dt} = \frac{k(v_{pl} - v_i) - b\frac{d\theta_i}{dt}}{a + \frac{G}{v_i}} + \frac{\theta_i}{2\beta\sigma_n},
\]

which applies for each surface within the shear zone. Our approach for including radiation damping is similar to that described in previous works (Perfettini & Avouac 2004, Ziv 2007).

We assume that the model begins with steady creep, and thus each surface of the fault zone undergoes steady state slip at velocity \(v_i = v_o\) with \(\mu_o = 0.6\) and \(\theta_{ss} = L/v_o\). The effective stiffness \(k_i\) between the load point and surface \(i\) within the fault zone is given by:

\[
\frac{1}{k_i} = \frac{1}{K_{ext}} + \sum_{j=1}^{i} \frac{1}{K_{int,j}}.
\]

To determine shear motion within the fault zone, we solve the coupled Equations 2, 4-5, using a \(4^{th}\) order Runga-Kutta numerical scheme. As noted above, perturbations in slip velocity are imposed at the shear zone boundary. This is assumed to occur via a remote loading stiffness \(K_{ext}\). Then, for each time step in the calculation, the surface with the lowest frictional strength is allowed to slip.

Our initial conditions are that shear and normal stress are the same on each surface. We ensure that time steps are small compared to the ratio of slip surface separation, \(h\), to elastic wave speed. Thus, within a given time step, only one surface slips and it is coupled elastically to the remote loading velocity via the spring stiffness given in Equation 5.

3.2 Frictional Response to Changes in Imposed Slip Rate

Figure 9 shows macroscopic shear strength of the fault zone as a function of slip at the fault zone boundary. Shear stresses are equal on all slip surfaces, however frictional strengths are not. Thus, Figure 9 shows friction of the weakest surface within the fault zone as a function of offset at the fault zone boundary. This case shows behavior of a fault zone that has homogeneous frictional properties.
Figure 9. Results from the model runs showing friction as a function of slip. For comparison, the intrinsic frictional response for a single surface is shown together with the frictional response of the shear zone. Note that the shear zone exhibits a prolonged phase of hardening, associated with the rate/state friction response of each layer, followed by weakening. See Table 1 for parameter values.

The macroscopic frictional response of the fault zone differs from the constitutive response of the individual surfaces within it. In particular, the fault zone exhibits a protracted phase of strain hardening prior to reaching the maximum yield strength (Fig. 9). The peak strength is reached in a slip displacement of $< 5\%$ of $D_c$ for a single surface, whereas the fault as a whole requires slip equal to $200\%$ of $D_c$ before weakening begins. As a result, the effective critical friction distance for the fault zone significantly exceeds that for an individual slip surface (Fig. 9). The maximum yield strength of the fault zone, which is proportional to the friction parameter $a$, is nearly identical to that for an individual surface. Finally, the steady-state frictional strength is the same in both cases (Fig. 9).

The relationship between the intrinsic frictional behavior of a surface and the zone of active shear, as a whole, is important for several aspects of earthquake rupture and shear localization. Figure 10 shows this relationship for a series of model runs to different shear strains. In each case the intrinsic frictional response of a single surface is shown versus slip on that surface. In addition, the frictional strength for the shear zone is plotted versus boundary slip. The two curves are plotted on the same scale; but note that the single surface is subject to larger total slip displacement, so as to illustrate the complete behavior. The panels of Figure 10 show three different amounts of shear applied at the boundary and below each plot is the spatial distribution of slip across the full shear zone.
Figure 10. Three snap shots of the relation between frictional behavior and slip distribution within a model fault zone. In each case the response of a single surface is plotted together with the behavior for the complete shear zone. The single surface is the same in each panel. The shear zone response in panels a-c is shown for progressively greater boundary shear. Images below each plot show slip distribution, via offset of markers that were initially vertical in this orientation. Note that strain is initially pervasive but that localization occurs abruptly during frictional weakening. See Table 1 for parameter values.
The frictional model indicates that a perturbation in slip rate at the shear zone boundary results first in pervasive shear, up to a point, followed by localization along a single surface (Fig. 10). Comparison of the friction curves and slip distribution shows that localization occurs at the point that frictional weakening begins. The initial period of hardening, dictated by the friction rate parameter $a$, is prolonged in the shear zone, compared to a single surface, because each surface must proceed through this hardening phase before the zone as a whole can weaken.

4 DISCUSSION

4.1 Shear localization and frictional behavior of granular layers

Our observations indicate that the critical slip distance for friction of granular materials represents the combined effect of multiple particle-particle contact interfaces. This is evident in the laboratory data on shear localization (Fig. 2) showing that dilation is confined to a fraction of the layer once shear becomes localized. Laboratory data for granular layers also show evidence of localization in the form of the critical slip distance for friction $D_c$ (Fig. 5). Our data show that laboratory measurements of $D_c$ represent the effective critical slip distance for the zone of active shear, which points the way toward a model for upscaling laboratory results to tectonic faults. Indeed, these laboratory data are one of the motivations for the numerical model presented here.

4.2 Spatio-temporal complexity of shear localization and delocalization

One of the enduring puzzles of shear localization in granular fault zones is that of shear band migration and delocalization. A typical fault zone in nature, and in the laboratory, includes multiple zones of shear localization, rather than a single zone. This may indicate a progressive process of localization, where one type of feature is active for a limited time and then another takes over. Or, it may indicate that a set of shear zones operate simultaneously, to produce a penetrative shear fabric. However, in either case, the existence of multiple slip surfaces raises a fundamental question: why does shear concentrate in one location and then switch locations? Is there a strain hardening process that begins once a shear band forms and, if so, does each shear band accommodate the same critical strain prior to abandonment? Another possibility is that strain localization is a local process within the bulk, and a given shear band minimizes the rate of work for only a confined region. In this case, it is important to know what sets the length scale of this region.

Our experiments involve simultaneous shear and comminution of granular materials. Previous works have documented the relationship between comminution and strain under conditions similar to ours (e.g., Mandl et al. 1977, Marone & Scholz 1989). Although our experiments include detailed measurements of macroscopic stress and layer strain, these data are of limited value in answering the most important questions raised above. One would like to have independent assessment of the spatial distribution of shear strain as a function of imposed shear on the layer. While this is beyond the scope of the present data, our measurements of layer dilation as a proxy for shear localization offer some information, and the numerical model provides some insight about how the intrinsic frictional response of surfaces within a zone effect the overall response of a fault zone.
5 CONCLUSIONS

Laboratory friction experiments combined with constitutive modeling provide a powerful means of investigating problems in shear localization. Our laboratory data show that layer dilation, in response to small perturbations in creep stress rate or strain rate, can be used as a sensitive proxy for the degree of strain localization. For granular layers sheared at normal stresses up to a few MPa, shear strain becomes fully localized prior to engineering shear strains of 1. Thin section analysis shows that shear localization is a progressive, rather than abrupt, process within a granular layer. Our experiments included velocity step tests, which probe the friction constitutive behavior and its relation to shear localization. We present a numerical model for shear within fault zones composed of multiple slip surfaces and use the model to evaluate shear localization. The spatial distribution of fault zone shear depends strongly on the intrinsic frictional properties of the materials and on elasto-frictional interaction. The model shear distribution is strikingly similar to the slip distribution documented in thin sections from experiment. Frictional processes determine the onset of strain localization and the width of active shear strain in granular shear zones. Our work has important implications for a range of conditions relevant to earthquake faulting and landslides.

ACKNOWLEDGMENTS

We thank Y. H. Hatzor and the other organizers of the Batshiva de Rothschild seminar on shear physics at the meso-scale in earthquake and landslide mechanics. The workshop was extremely stimulating and very enjoyable. We gratefully acknowledge support from the National Science Foundation under grant numbers ANT-0538195, EAR-0510182, and OCE-064833. J. Samuelson, A. Niemeijer, and B. Carpenter are thanked for stimulating discussions during the course of this work.

REFERENCES


Appendix B

This appendix represents work in which I am a second author, but in which I conducted the experiments for and contributed many of the ideas. This work considers the energy budget of earthquakes and is related to many aspects of my thesis, especially Chapter 4. Future work is planned combining the methods of this Appendix of measuring temperature increase and grain size reduction and the acoustic emission to look at seismic efficiency, as is possible with the data in Chapter 4. The following manuscript will be submitted to Earth and Planetary Science Letters.

Experimental constraints on energy partitioning during stick-slip and stable sliding within analog fault gouge

**Patrick M. Fulton**\(^1,2\)

**Andrew P. Rathbun**\(^1\)

\(^1\)Department of Geosciences, Pennsylvania State University, University Park, PA 16802

\(^2\)College of Oceanic and Atmospheric Sciences, Oregon State University, Corvallis, OR 97331

**Abstract**

The lack of substantial frictional heat anomalies across major fault zones has been a key observation suggesting that faults support low shear stress during slip. Some studies have suggested that the lack of large heat anomalies may be a result of considerably less energy going to frictional heat than generally thought and that a large fraction of energy is dissipated by other processes such as the creation of new surface area. We evaluate this hypothesis through the analysis of 19 laboratory shear
experiments for both stick-slip (seismic) and stably sliding (aseismic) analog fault gouges. These experiments differ from previous laboratory studies in that they 1) provide independent constraints on frictional heat generation and energy consumed generating new surface area, 2) cover a broader range of shear stresses than most previous studies (2 – 20 MPa), and 3) evaluate both stick-slip and stable sliding within granular material. Based on the analysis of high-precision temperature measurements and comparisons with numerical model simulations, we show that >90% of the total energy goes to frictional heat generation ($E_H$) for all of our experiments, and based on grain size analysis that ~1% of total work is consumed generating new surface area ($E_{SA}$). These results are consistent with assumptions allowing frictional resistance to be inferred from thermal data and suggest there is no relationship between stick-slip or stable sliding within fault gouge and large fractions of total energy going to new surface area generation and/or small fractions to frictional heat.

1 Introduction

Thermal data have played an important role in evaluating the mechanics of earthquakes and faulting. The lack of large frictional heat anomalies across major fault zones in regional heat flow data or in borehole temperature profiles that intersect faults after a large earthquake has been one of the primary observations suggesting that many faults support low shear stress during slip, considerably less than expected by laboratory-derived friction laws and hydrostatic pore pressure [e.g., Brune et al., 1969; Lachenbruch and Sass, 1980; Wang et al., 1995; Kano et al., 2006; Tanaka et
An important assumption that allows for frictional resistance during slip to be inferred from thermal observations is that nearly all of the dissipated energy during fault slip goes to frictional heat generation [e.g., Brune et al., 1969; Lachenbruch and Sass, 1980]. Figure 1 illustrates how work during slip is partitioned to elastic radiation (e.g., seismic waves) and dissipated energy.

Total work during slip is defined by the sum of the work due to shear and the sum of work due to slip-induced dilation or compaction. This is expressed by equation 1,

$$ W = A \int_0^D \tau d\delta + A \int_0^L \sigma_n dw = A \tau D + A \sigma_n L $$

**Equation 1,**

where $A$ is the fault surface area, $\tau$ is the displacement-averaged shear stress in the direction of slip, $D$ is the total displacement, and $L$ is the change in thickness due to compaction or dilation during slip. The total work during slip is balanced by dissipated energy $E_f$ and radiated energy $E_a$. Elastic radiated energy $E_a$ is related to the stress drop during stick-slip unstable sliding and the apparent stress $\tau_a$, which can be determined seismologically. It is generally considered to account for $<6\%$ of the total work during slip [McGarr, 1999], whereas dissipated energy $E_f$ is thought to account for $\sim 95\%$ of the total work [e.g., Lachenbruch and Sass, 1980; Lockner and Okubo, 1983]. Dissipated energy includes both frictional heat and fracture energy, which can include work done by chemical processes, dilation, and grain rolling, in addition to the energy consumed making new surface area through rock fracture and grain breakage. Dissipated energy $E_f$ is a function of the displacement-averaged frictional resistance along the fault during slip $\tau_f$ and cannot be directly determined
seismologically, although it is a critical parameter in controlling the mechanics of fault slip.

Some studies have argued that the lack of thermal anomalies across major fault zones may not necessarily imply that the average frictional resistance and shear stress during slip has been low, but rather that the fraction of dissipated energy going to frictional heat (i.e., the thermal efficiency) may be considerably less than ~90 - 95% of the total work, and that the missing heat energy is partitioned to other processes [e.g., Brown et al., 1998; Wilson et al., 2005]. Understanding how energy during slip is partitioned between frictional heat, seismic radiation, and the generation of new surface area and other processes is important not only for characterizing fault strength, but also for our general understanding of the mechanics of earthquakes and faulting and for assessing seismic hazard.

Here we investigate how energy is partitioned during slip through the use of laboratory experiments of shear within analog fault gouge material. Although laboratory experiments may be an effective way to put constraints on the energy budget of fault slip, few experimental studies within the geosciences literature have tried to directly constrain the amount of frictional heat generation during slip [Lockner & Okubo, 1983; Yoshioka, 1985; Blanpied et al., 1998; Brown, 1998; Mair & Marone, 2000] or the amount of energy consumed in generating new surface area [e.g., Engelder et al., 1975; Yoshioka, 1986]. The few experimental studies of thermal efficiency (i.e. the fraction of total work during slip that is spent generating frictional heat) have generally been performed on granite slabs and the detrital material generated during the experiment [Lockner and Okubo, 1983; Yoshioka, 1985;
Brown, 1998] or on stably sliding granular material [Lockner and Okubo, 1983; Mair and Marone, 2000]. The results of these experiments have generally supported estimates of > 90 - 95% of the total work during slip going to frictional heat generation.

Some observations, however, have raised questions regarding our understanding of the earthquake energy budget [e.g., Yoshioka, 1985; Brown, 1998; Mora and Place, 1998; Wilson et al., 2005]. For example, the grain size distribution of some natural fault gouges have been interpreted to suggest that the creation of new surface area through grain breakage may amount to as much as 50% or more of the total work during slip rather than ~ 1% as is more commonly thought [Wilson et al., 2005]. This interpretation, although controversial [e.g., Chester et al., 2005; Rockwell et al., 2009], has been used to suggest that considerably less energy goes to frictional heat generation and may thus explain the lack of large frictional heat anomalies across major fault zones without the need for low shear stress during slip.

Some experimental results of frictional heat generation during stick-slip (earthquake-like) sliding of granite slabs have also suggested that thermal efficiency may be less than conventionally thought [Yoshioka, 1985; Brown, 1998]. The experimental results of Brown [1998] reveal a significantly large difference in the rate of temperature rise, a proxy for frictional heat generation rate, between stick-slip and stable sliding, suggesting a thermal efficiency ~50% for stick-slip failure rather than >90% interpreted for stable sliding under similar stress conditions. Brown [1998] argues that the discrepancy between stick-slip and stable sliding systems is not a result of pulsed versus continuous heat generation, rate and state friction, or thermal
pressurization. It was also presumed that the generation of new surface area through grain size reduction was negligible, based on the small amount of detrital material between the layers after each experiment, although it was not directly verified through analysis. The abnormal thermal efficiency in these experiments is interpreted to only occur at normal stresses > ~7 MPa, and thus would explain why this behavior was not seen in similar experiments by Lockner and Okuba [1983] which were conducted at normal stresses < 3.45 MPa. Similar interpretations of low thermal efficiency have also been determined within laboratory experiments of slip between granite slabs that exhibit chaotic stick-slip behavior and stress drops that are very large in both total magnitude (~20 MPa) and in relation to the average background stress [Yoshioka, 1985]. These results appear to be largely a function of an absence of abundant gouge / detrital material within the slip zone during the experiment. Large earthquakes, however, are generally hosted within mature fault zones that have well-established gouge zones that support slip [e.g., Scholz, 2002]. Results of numerical models of shear within granular gouge material have suggested that grain interactions including bouncing and rolling of grains may have a significant influence in reducing thermal efficiency [Mora and Place, 1998], although the models do not include the effects of grain size reduction in either consuming energy or restricting rolling of grains. Experiments of shear heating for stably sliding (aseismic) gouge material do not reveal low thermal efficiency [Mair and Marone, 2000].

Although interpretations of low thermal efficiency and/or large fractions of energy consumed by grain breakage are unusual, the examples described above illustrate a level of uncertainty in our understanding of the partitioning of energy
during fault slip, particularly that associated with frictional heat generation and the generation of new surface area. These studies raise the question of whether the creation of new surface area may be a considerably larger fraction of total energy during stick-slip sliding than generally considered and whether this or similar processes contribute toward significantly reducing the thermal efficiency of earthquakes to levels ~ 50% or less, as some have hypothesized [e.g., Brown, 1998; Wilson et al., 2005].

Here, we evaluate the partitioning of energy during slip through the analysis of 19 laboratory shear experiments of both stick-slip (seismic) and stable (aseismic) sliding within analog fault gouge (Table 1). These experiments are particularly relevant in that 1) they were designed such that constraints on both the amount of frictional heat generation and energy consumed in making new surface area can be independently determined, 2) the sliding behavior for each material used is consistently similar between experiments, 3) they cover a greater range and magnitude of stress conditions than most previous experiments of frictional heat generation during stick-slip sliding, and 4) they cover both stick-slip and stable styles of sliding within analog fault gouge, whereas previous laboratory studies of thermal efficiency have been performed on granite slabs and the detrital material generated during the experiment or on only stably sliding granular material [Lockner and Okubo, 1983; Yoshioka, 1985; Brown, 1998; Mair and Marone, 2000].

The objective of this study is to test the hypothesis that the generation of new surface area through grain breakage during stick-slip sliding within fault gouge accounts for a significantly large portion of the total energy budget during slip and
thus reduces the amount of energy partitioned to frictional heat. Fundamental questions we seek to address include: 1) Does frictional heat consistently account for ~ 90% or more of the total slip energy budget for a range of different stress conditions? 2) Is the generation of new surface area through grain breakage a large, yet overlooked, component of the energy budget? 3) Is there a significant difference in the partitioning of energy to frictional heat or to new surface area depending on the mechanism of slip (stick-slip vs. stable sliding)?

In the following sections we describe our experimental setup and discuss the data and related analysis that quantifies the amount of energy associated with stress drops, creation of new surface area, and frictional heat for each experiment. We then discuss how our findings address the questions above, relate to other studies of slip energy budgets, and may help inform our understanding of earthquake mechanics.

2. Experimental Setup

Our experiments were conducted at the Penn State Rock and Sediment Mechanics Laboratory in a servo-controlled, double-direct shear apparatus (Figure 2A, inset) at normal stresses ranging from 5 to 50 MPa and a driving velocity of 200 μm/s for all experiments except p1901 which was run with a driving velocity of 100 μm/s in order to evaluate any effect on the stick-slip behavior. This velocity range produced stress drops during stick-slip sliding experiments similar to those determined in natural events [Allman and Shearer, 2009]. For the purposes of this study, we do not attempt to further explore the range of driving velocities and stress drop size. Table 1 lists many of the relevant details for each experiment. In order to evaluate differences
between stick-slip and stable sliding, experiments were conducted on two separate materials with similar initial grain size distribution uniformly around 120 µm. Soda lime glass beads which exhibit stick-slip sliding behavior were obtained from the Mo-Sci Corporation, Rolla, Missouri and stably sliding Ottawa sand of >99% pure angular quartz was obtained from US Silica Company, Ottawa, Illinois. By using two granular materials with similar initial grain size, but different sliding stability, we are able to evaluate both stick-slip and stable sliding for similar stress conditions and loading rates without having to make ad hoc changes to the shear apparatus and system stiffness [e.g., Brown, 1998]. For each experiment two granular layers, 2 mm thick were sandwiched within a three-block arrangement in a manner similar to the experiments of Mair and Marone [2000] (Figure 2A, inset). The side blocks were held stationary and the center block was driven downward at a constant rate, causing shear. Nominal contact area was held constant at 10 cm by 10 cm with each steel block grooved perpendicular to the shear direction to ensure deformation happens within the gouge zone, not at the steel-gouge interface. Most experiments were run for ~100-140 seconds for total displacements of > 20 mm (Table 1), corresponding to shear strain values >10. Repeat experiments were conducted with shorter displacements so that first-order differences in energy partitioning as a function of total displacement and grain breakage could be assessed.

To measure the thermal response of frictional heating during the experiments four T-type thermocouples were placed inside one steel side block 3 mm, 4 mm, and 10 mm from the shear zone and with a redundant measurement at 3 mm. With the use of amplifier circuits the thermocouples have an effective measurement precision of ~ 0.1
°C throughout the experiments (Figure 2A). Thermocouples were calibrated at 0 and 24 °C. Values of shear stress, normal stress, layer thickness, shear displacement and temperature were recorded at a constant rate of 1 kHz or greater throughout each experiment. The resulting data allow for the calculation of total work based on Equations 1 and 2 and the summing of work during displacement by stick-slip events and the work during displacement by creep. To account for apparatus effects we remove displacement associated with the elastic stretching from the measured load point displacement of the horizontal and vertical rams. Measured elastic stiffnesses are 5 and 3.7 MN/cm for the vertical and horizontal rams, respectively. Displacement and work during stick-slip experiments are mostly accumulated through stick-slip (stress-drop) events, whereas stable sliding experiments they are entirely accumulated through creep.

3 Experimental Results and Analysis

3.1 Energy associated with stress drops

Figure 2 shows results typical of one of our stick-slip sliding experiments. The vertical width of the shear stress vs. displacement curve in Figure 2A and panels A and B of Figure 3, reflects the magnitude of the stress drops associated with several hundred stick-slip events recorded during each stick-slip sliding experiment. A more detailed view of data from a typical series of stick-slip events is shown in Figure 2B. The high-resolution of the measurements allows us to calculate the amount of energy released by the stress drop, $E_{st}$.
\[ E_{\Delta \tau} = \frac{\Delta \tau u A}{2} \]  \hspace{1cm} \text{Equation 3,}

where \( \Delta \tau \) is the stress drop, \( u \) is the displacement during the stress drop and \( A \) is the rupture area which we assume to be equal to the experimental contact area. This energy is illustrated by triangle ABC in Figure 1. Assuming a simple slip-weakening model [e.g., Andrews, 1976] this energy would consist of a combination of seismic radiation along with fracture energy, which is a combination of work done from grain interactions, dilation, and the creation of new surface area. For more complex slip weakening this estimate may not account for all the fracture energy during slip (Figure 1) [Kanamori and Rivera, 2006]. Calculation of \( E_{\Delta \tau} \) allows for the fraction of total energy associated with stress drops and presumably changes in overall energy partitioning to be monitored over the duration of each experiment. The total work during the experiments are calculated by Equation 1 where we calculate the work done from displacement during stick-slip events as well as work done during aseismic creep displacement.

Figure 3C shows results of these calculations for stick-slip sliding experiments at two different applied normal stresses, p1553 with 40 MPa (blue data) and p1826 with 7.5 MPa (red data). The results of p1826 are typical of our stick-slip experiments with normal stresses < 15 MPa; estimates of the energy associated with the stress drop for each stick-slip event are consistently \( \sim 6\% \) of the total coseismic work. In contrast, higher stress experiments such as p1553 show a substantial change in values over the course of the experiment, with the fraction of total coseismic work associated with the stress drop reducing to levels considerably less than in the lower
stress experiments later in the experiment. The values are still consistently around 6% and less.

The changes in energy associated with stress drops during the higher stress experiments also correspond to changes in shear stress. **Figure 3A** shows how after the initial runup of shear stress for p1553, shear stress increases until ~50 s into the experiment while the fraction of total energy associated with the stress drops decreases due to both an increase in shear stress leading to larger total work and smaller magnitude stress drops. In contrast, lower stress experiments such as p1828 show roughly no change in shear stress or the fraction of work associated with stress drops after the initial shear stress runup (**Figure 3B**). We interpret the changes in the higher stress experiments as a change in energy partitioning which likely results from large amounts of grain breakage during the beginning of the high stress experiments and reduces after a certain degree of grain size reduction. As grains fracture, the overall grain size distribution and angularity increases, both of which promote stable sliding, leading to an increase in friction and shear strength, as well as smaller stress drops [e.g., *Mair et al.*, 2002; *Anthony and Marone*, 2005].

**Figure 4** illustrates that the amount of work done by co-seismic slip and dilation considerably decreases at ~50 s and that stable preseismic creep begins to account for an increasing fraction of the total work consistent with this interpretation. For the lower stress experiments, such as p1828, the different sources of work remain at a similar fraction throughout the experiment. For p1828 with normal stress of 7.5 MPa, the amount of work from preseismic slip is comparable to coseismic slip. Amongst
the stick-slip experiments the fraction of work done by preseismic slip decreases as a function of increasing normal stress.

3.2 Energy Consumed in Generating New Surface Area

To determine the amount of energy partitioned to the creation of new surface area through grain size reduction, we conducted grain size analysis on post-shear material from each experiment using a Malvern Mastersizer 2000 laser diffraction grain size analyzer. Following standard practices, samples were sonicated to disaggregate the particles prior to placing them into the Mastersizer and then were kept in the machine for <1-2 minutes in order to avoid settling [e.g., Rockwell et al., 2009]. Figure 5A shows an example of the grain size distribution of both pre- and post-shear material.

The energy consumed in creating new surface area $E_{SA}$ is computed using Equation 4 [e.g., Wilson et al., 2005; Chester et al., 2005; Ma et al., 2006; Rockwell et al., 2009],

$$E_{SA} = \gamma \Delta S \quad \text{Equation 4},$$

where $\gamma$ is the material specific surface energy (J/m$^2$) and $\Delta S$ is the difference in surface area (m$^2$) between pre- and post-shear material. Quartz and soda-lime glass have specific surface energies of 1 and 4 J/m$^2$, respectively [Iler, 1979; Wiederhorn, 1969]. The surface area $S$ of both pre- and post-shear sample material is computed from grain size distribution data by:
\[ S = R \sum_{i} \frac{6}{d_i} \%V_i V \]  
**Equation 5,**

where \( V \) is the total volume of the sample used during the shear experiments, \( \approx 1 \times 10^{-5} \text{ m}^3 \), \( d \) and \( \%V \) is the grain diameter and volume percent of for each grain size bin \( i \), and \( R \) is a roughness factor. Grain size distribution is measured in bins for a range of grain size diameters that increase stepwise by a factor of 1.165 from \( 5.82 \times 10^{-8} \text{ m}^2 \) to \( 8.79 \times 10^{-3} \text{ m}^2 \). Both of our sample materials have relatively uniform initial grain size of \( \approx 1.2 \times 10^{-4} \text{ m}^2 \) (**Figure 5A**). The change in surface area from the grain size distribution data is calculated assuming the shape of each grain is spherical and multiplying by an average grain roughness factor, \( R \) [e.g., Wilson et al., 2005; Chester et al., 2005; Ma et al., 2006]. Grain roughness factor is poorly constrained for pulverized material and is often assumed based on values determined for mechanically ground material or natural field samples, which may be greatly affected by the effects of weathering [e.g., Hochella and Banfield, 1995]. Previous studies that have determined energy estimates from grain size distributions of natural fault gouge have either determined or assumed values \( \approx 5 \) for grain roughness factor [e.g., Wilson et al., 2005; Chester et al., 2005; Ma et al., 2006]. Here, we report the results without multiplying by a grain roughness factor because our starting materials start as nearly-spherical grains and the degree of grain breakage and roughening is not consistent between experiments under different stress conditions. However, for one of the larger normal stress strike-slip experiments, p1553, an independent measure of post-shear surface area was determined by Barrett-Emmett-Teller (BET) \( \text{N}_2 \) adsorption analysis [Gregg and Sing, 1982; Hochella and Banfield, 1995].
Comparison of the results of this analysis with the estimate of surface area from grain size analysis suggests a maximum grain roughness factor of ~3 for our experiments. This result is consistent with other grain size roughness determinations for post-shear material from laboratory experiments of stick-slip sliding [Yoshioka, 1986].

Estimates of the energy consumed by creating new surface area through grain size reduction determined from grain size analysis, $E_{sa}$, are illustrated in Figure 5B as the fraction of total work against the average shear stress for each experiment. All of the results are < 1% of the total work except experiment p1584, which was a repeat experiment of p1553 performed at the same applied normal stress, but with a shorter total displacement and slightly smaller average shear stress (Table 1). Although the results of p1553 and p1584 with similar stress conditions, but different total displacements support the interpretation that grain size reduction is more significant earlier in the experiments, because the shorter displacement experiment appears to contribute a slightly larger fraction of energy to new surface area, this result is likely a reflection of p1584 being an outlier within the range of uncertainty in the method of determining $E_{sa}$ from grain size analysis. Experiment p1562, which is also a high stress, low-displacement experiment, does not appear to have a percent work to new surface area estimate much different than the other experiments.

Changes in the amount of energy consumed by grain breakage, however, likely do exist as a function of displacement during our experiments; as grain size distribution widens, smaller particles work to buffer large ones and create complex force chains that distribute stress leading to less fracture and probably lead to the smaller stress drops noted above in our higher stress stick-slip experiments [e.g., Anthony et al.,]
2005; Sammis and Ben-Zion, 2008]. These changes, if present, appear to be beyond the resolution of our analysis here, largely due to the likely small magnitude of these changes and the small fraction of total work the process appears to consume overall.

Based on the grain roughness estimate described above, our estimates of the percent work consumed by the creation of new surface area (Figure 5B) may be scaled by as much as a factor of 3. Even with this multiplication factor, these estimates would still be consistent with most previous analyses of laboratory experiments, natural fault gouge, and numerical models that have determined values ~1% [e.g. Chester et al., 2005; Shi et al., 2008; Rockwell et al., 2009]. In order to obtain values as high as 50% as interpreted for fault gouge sampled from a freshly formed fault [Wilson et al., 2005], our particle size distribution would require a considerably large volume percent of material with particle diameters ~10^{-1} \mu m or smaller for experiments such as p1553. Our results do not reveal much in terms of volume fraction of grains this small for any of our experiments (e.g., Figure 5A).

3.3 Frictional Heat Generation

Unlike most previous experiments that have compared thermal efficiency between stick-slip and stable sliding of granite slabs [Brown, 1998], our experiments are performed on granular gouge and cover a broad range of different stress conditions. Figure 6 shows the rate of initial temperature rise determined for each of our stick-slip and stable sliding experiments for the range of average shear stress conditions covered. The results show a linear trend in the rate of initial temperature rise as a function of average shear stress. In addition, they do not reveal any
significant difference between stick-slip and stable sliding for similar shear stress conditions and loading velocity (Figure 6). These results suggest that experiments of stable sliding and stick slip motion consistently have similar thermal efficiency as a function of stress and for both styles of sliding.

To constrain the values of thermal efficiency during our experiments, we compare the observed temperature signal to numerical model simulations of frictional heating based on the shear stress and displacement data for each experiment. The model consists of a 3D model domain representative of our three-block experiment arrangement and contains thermal conductivity and diffusivity values appropriate for the steel blocks and sample layers. Room temperature is set as a boundary condition along each external face and frictional heating is prescribed within two 1 mm thick layers, representative of our sample material undergoing shear. The volumetric frictional heat generation rate is calculated based on the shear stress and loading rate (constant) measured during each experiment and is updated within the model simulations every one second of simulation time. The model simulations are run for different thermal efficiency values from 10% of the total work going to frictional heat to 100%. Based on comparisons between model results and observed temperatures during stable sliding at a range of normal stresses, the model results appear to give consistent and reasonable results. We do not attempt to resolve the thermal efficiency to values less than 10% for several reasons – the models are simplifications of the shear heating process and do not take into account material transport during shearing, heterogeneity in frictional heat sources or strain localization less than 1 mm, and in addition, the model does not account for the possibility of changes in thermal
efficiency during the experiment. The models, however, appear sufficiently adequate for the purpose of testing whether frictional heat generation accounts for ~90% or more of the total slip energy or ~50%.

For both stick-slip and stable sliding a comparison between simulated temperatures and the observed temperatures within the side block consistently support the interpretation of ~90% or more of the total work during slip going to frictional heat generation as opposed to ~50% (e.g. Figure 6 inset). In the higher stress experiments, such as p1553, a subtle change in the temperature rise pattern that deviates from the pattern predicted by the models is apparent at ~50 s (Figure 2 and Figure 6 inset). It is unclear whether this pattern reflects complexities not accounted for in the model or rather a small (<10%) change in thermal efficiency during the experiment. This transition also appears to correspond to the cessation in the rise of shear stress associated with grain fracture (e.g., Figure 3) and changes in energy associated with stress drops, which encourages us to make the latter interpretation suggesting a subtle change in thermal efficiency.

4. Discussion and Conclusions

The objective of this study was to determine whether frictional heat consistently accounted for ~90% of the total slip energy during both stick-slip and stable sliding of granular material analogous to fault gouge for a range of stress conditions and whether the energy consumed in making new surface area through grain breakage was a significantly larger component of the energy budget then generally thought. We find roughly equivalent rates of temperature rise between stick-slip and stable
sliding for all stress conditions evaluated, from an average initial shear stress of 2 to
21 MPa, implying that there is not a considerable difference in thermal efficiency
between styles of sliding and no significant change as a function of average shear
stress. In addition, comparison of the observed temperature signals during
experiments with numerical models of heat transfer based on the stress and
displacement data suggest the observed heat signals for both sliding styles are most
consistent with thermal efficiency values ~90% or greater.

This determination of large thermal efficiency during stick-slip sliding is
consistent with analysis of exhumed pseudotachylite-bearing fault rocks and
numerical simulations of fault rupture processes [Pittarello et al., 2008; Shi et al.,
2008]. Similar values have also been indirectly inferred based on constraints for
other significant components to the earthquake energy budget that suggest <=6% of
the slip energy goes toward seismic radiation [McGarr, 1999] and <=1% to the
creation of new surface area through grain breakage [Engelder et al., 1975; Yoshioka,
1986; Chester et al., 2005; Rockwell et al., 2009].

The results of our analysis also suggest that the generation of new surface area
through grain breakage does not consume a significantly large fraction of the total
slip energy. Estimates of new surface area, determined by both grain size analysis
and gas absorption methods correspond to ~ 1% of the total work due to slip. We
find only a small fraction of total energy is being consumed in making new surface
area, which is consistent with previous determinations based on both laboratory and
field data [e.g., Engelder et al., 1975; Yoshioka, 1986; Chester et al., 2005; Rockwell
et al., 2009], as well as estimates of fracture energy inferred from models of ground
motion waveforms [Tinti et al., 2005] and numerical models of rupture mechanics [Shi et al., 2009]. These results, however, are in marked contrast to the interpretations of fault gouge by Wilson et al. [2005] that suggest values ~50% of the total slip energy. One possible explanation for the discrepancy with Wilson et al.’s unusually large estimate is that this value was determined based on gouge from the 1997 M 3.7 Bosman earthquake that ruptured through intact rock creating a new fault zone. In the present study, we have focused on the energetics of slip within fault gouge, similar to how we expect most large earthquakes to be supported by existing fault zones within established gouge zones.

Overall, our interpretations of energy partitioning during stick-slip sliding are consistent between each independently constrained quantity of the energy budget. In our experiments >90% of the total energy appears to go to frictional heat generation ($E_H$), ~1% to generating new surface area ($E_{SA}$), and ~4% associated with the stress drop ($E_{\Delta\tau}$), which may include energy that contributes to grain breakage in addition to the elastic radiation. The results suggest that for these experiments nearly all of the dissipated energy during slip (i.e., total work minus elastic radiation) goes to frictional heat generation. This supports the assumption used in interpreting thermal data along major active fault zones in terms of frictional resistance during slip, although we note that directly relating our laboratory results to earthquake mechanics is not trivial since our experiments do not include many temperature-related processes including co-seismic chemical alteration and dynamic weakening due to thermal pressurization that may be active during natural earthquakes [e.g., Andrews, 2002; Hamada and Hirona, 2010]. Our results do, however, suggest that there is not a
fundamental relationship between stick-slip or stable sliding within fault gouge and large fractions of energy consumed in the generation of new surface area and/or low thermal efficiency.

Acknowledgements

Grain size analysis and N\textsubscript{2} BET measurements were conducted at the Materials Characterization Lab, Penn State University.

References


Rockwell, T., M. Sisk, G. Girty, O. Dor, N. Wechsler, and Y. Ben-Zion (2009), Chemical and physical characteristics of pulverized Tejon Lookout Granite


Figure 1: The total work per unit area of fault during a stick-slip event. The total work is defined by Equation 1 and represented by the total area shown in the figure. The bold line represents the shear stress felt by the fault during slip based on an idealized simple slip-weakening model [e.g., Andrews, 1976]. The dashed curve illustrates how this slip-weakening may be considerably more complex. The area beneath the slip-weakening curve represents the dissipated energy which is made of frictional heat $E_H$, and fracture energy $E_G$ consisting of energy making new surface area $E_{SA}$ and may also include additional energy consumed by dilation, grain rolling and other grain interactions. The difference in total and dissipated energy is released as elastic radiation $E_R$. 
Figure 2: Representative data measured during a high normal stress shear experiment. Panel A illustrates the three-block double direct shear arrangement and a temperature measurement from a thermocouple 3 mm from the sliding surface and shear stress for a stick-slip sliding experiment. The width of the shear stress curve illustrates the magnitude of stress drops during stick-slip events. The arrow points to the location of subsample of typical stick-slip events shown in detail in Panel B. The elastic stretching of the apparatus is removed from the load point displacement curve to produce the fault displacement, as described in the text.
Figure 3: (A and B) Shear stress curves for typical high- and low-stress stick-slip experiments (described in the text). The higher stress experiments show an increase in stress and decrease in average stress drop likely due to increased grain size distribution and grain angularity. (C) Percent of work done during a stick-slip event associated with the stress drops ($E_n/W*100$) as a function of time during these two experiments.
**Figure 4:** The relative source of work as a percentage for each stick-slip event during a typical high- (panel A) and low-stress (panel B) stick-slip experiment, as defined in the text. For each event the total work includes the work due to co-seismic slip (closed blue circles) and dilation/compaction (green open circles) and pre-seismic creep (red x’s).
Figure 5: (A) Typical results of grain size analysis and on pre-shear (dashed curve) and post-shear (solid curve) sample material for our stick-sliding material. The stably sliding material has a similar initial pre-shear grain size distribution. (B) Percent of the total work during the experiment estimated to be consumed creating new surface area ($E_{SA}/W_{total} \times 100$) for each experiment versus the average shear stress. The estimates presented assume spherical grains or a roughness factor $R$ equal to 1. The scatter in data is likely a reflection of methodological uncertainty.
Figure 6: The rate of initial temperature rise is plotted against the average shear stress each calculated during a 10 second interval after the initial runup in shear stress for each experiment. Stick-slip experiments (red circles) and stable-sliding experiments (blue X’s) both show linear trends that do not significantly differ for each other suggesting consistent values of thermal efficiency for both styles of sliding at a range of shear stress conditions. The inset shows an example temperature record compared with numerical models of temperature rise. The models consistently suggest that 90% or more of the total work during each experiment goes to frictional heat ($E_H/W_{total}$*100) rather than smaller values such as 50%.
Table 1: Experiment Table. Stable sliding and stick-slip experiments are conducted on soda-lime glass beads and Ottawa sand, respectively.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Normal Stress (MPa)</th>
<th>Average Shear Stress (MPa)</th>
<th>Sliding Style</th>
<th>Displacement (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>p1492</td>
<td>10</td>
<td>3.1</td>
<td>Stable</td>
<td>28</td>
</tr>
<tr>
<td>p1493</td>
<td>5</td>
<td>6.6</td>
<td>Stable</td>
<td>28</td>
</tr>
<tr>
<td>p1510</td>
<td>15</td>
<td>5.2</td>
<td>Stick-slip</td>
<td>28</td>
</tr>
<tr>
<td>p1515</td>
<td>5</td>
<td>1.9</td>
<td>Stick-slip</td>
<td>22</td>
</tr>
<tr>
<td>p1516</td>
<td>10</td>
<td>3.8</td>
<td>Stick-slip</td>
<td>21</td>
</tr>
<tr>
<td>p1518</td>
<td>25</td>
<td>9.1</td>
<td>Stick-slip</td>
<td>21</td>
</tr>
<tr>
<td>p1552</td>
<td>10</td>
<td>3.8</td>
<td>Stick-slip</td>
<td>21</td>
</tr>
<tr>
<td>p1553</td>
<td>40</td>
<td>18.7</td>
<td>Stick-slip</td>
<td>21</td>
</tr>
<tr>
<td>p1562</td>
<td>50</td>
<td>15.4</td>
<td>Stick-slip</td>
<td>7</td>
</tr>
<tr>
<td>p1563</td>
<td>20</td>
<td>7.5</td>
<td>Stick-slip</td>
<td>21</td>
</tr>
<tr>
<td>p1564</td>
<td>27</td>
<td>15.7</td>
<td>Stable</td>
<td>21</td>
</tr>
<tr>
<td>p1580</td>
<td>20</td>
<td>12.1</td>
<td>Stable</td>
<td>21</td>
</tr>
<tr>
<td>p1581</td>
<td>30</td>
<td>13.5</td>
<td>Stick-slip</td>
<td>21</td>
</tr>
<tr>
<td>p1582</td>
<td>33.3</td>
<td>19.5</td>
<td>Stable</td>
<td>21</td>
</tr>
<tr>
<td>p1583</td>
<td>8.3</td>
<td>4.9</td>
<td>Stable</td>
<td>21</td>
</tr>
<tr>
<td>p1584</td>
<td>40</td>
<td>13.9</td>
<td>Stick-slip</td>
<td>8</td>
</tr>
<tr>
<td>p1826</td>
<td>7.5</td>
<td>2.8</td>
<td>Stick-slip</td>
<td>21</td>
</tr>
<tr>
<td>p1828</td>
<td>25</td>
<td>9.7</td>
<td>Stick-slip</td>
<td>21</td>
</tr>
<tr>
<td>p1829</td>
<td>16</td>
<td>9.9</td>
<td>Stable</td>
<td>21</td>
</tr>
</tbody>
</table>
Appendix C: Matlab codes

1. Stick slip picker

Code to pick stick slips. Input is a text file. The user updates terms in the first section and the rest of the code runs. The user defines and window size of how many points to consider at a time, and a tolerance or how big of a change in shear stress in that window. If the change exceeds the tolerance a stick slip is picked. The code outputs 9 plots, shown as Figures 1-9, and a text file. The given code is for experiment p1553.

```matlab
clear all
close all

%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%% % Code to pick stick slip and calc energy %%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%% 
A. Rathbun   June 16th 2008 various updates
Version June 24 2010
%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%% 

% User defined variables will change these for each experiment %

%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%% record=1000;    %Recording rate (Hz);
name='p1553';    %Experiment number
normstress = 40; %Normal Stres
velocity = 200;  %Velocity
experiment = load ('../../p1553.txt');
window=50;  % How many records to consider when looking for stickslips
tolerance=0.5; % Tolerance for picking stick-sips. This will take them if
              % stress drop is greater than 0.5 MPa
begin=8;       %Time to start looking at stick slips around
               %8 seconds into the experiment for p1553

plotter='y';   % Bool variables to plot stick slips  y to plot
               %if plotter = 'y' calc must = 'y'
saver='y';     % Save data file  y to save
calc='y';      % Calculate parameters  y to calculate
layertrend='n'; %Take out trend in layer?
area = 0.1 ;  %m2
```
rec = experiment(:,1); % Load data
lp_disp = experiment(:,2);
shear_stress = experiment(:,3);
layer = experiment(:,4);
%norm = experiment(:,5);
time = experiment(:,6);
friction = experiment(:,7);
%strain = experiment(:,8);
cec_disp = experiment(:,9);
%T1 = experiment(:,10); T2 = experiment(:,11);
%T3 = experiment(:,12); T4 = experiment(:,13);

clear experiment %Clear big variable for memory

posit = find(time<100); %Clear big variable for memory
othertime=time(posit);

count=1; % Counter for populating vectors each count will be one stick slip
finish=floor(length(time)/window)*window-window; %Find the end of the data
start=begin*record; %offset to the portion you want to start looking at.

for i=start:window:finish; %Loop to find stick slips
   [temptop,temprec1]=max(shear_stress(i:i+window)); %[max/min, position]
   [tempbottom,temprec2]=min(shear_stress(i:i+window)); % Looks at shear
   %stress and takes the max and min over the window. Stores these in
   %temp variables. Will be stored if max-min exceeds the user tolerance
   diff=temptop-tempbottom;
   if diff > tolerance % Store data if the condition is meet
      b(count)=temprec1; %Store the position of the max
   end
end
\begin{verbatim}
\begin{verbatim}
\texttt{d(count)=temprec2; \%min}
\texttt{b(count)=b(count)+i-1; \%Change to real position not just in each}
\texttt{d(count)=d(count)+i-1; \%iteration of i, a dummy variable.}
\texttt{avestress(count)=mean(shear\_stress(i:i+window));}
\texttt{count=count+1; \%Update counter, counts each stickslip, starts at 1.}
\texttt{end}
\texttt{end}

\texttt{if layertrend == 'y'}
\texttt{a=polyfit(othertime, layer(1:length(othertime)),1);}
\texttt{d_y = polyval(a,othertime);}
\texttt{aver=mean(d_y);}
\texttt{for count = 1:length(othertime);
\texttt{layer(count)=layer(count)-d_y(count)+aver;}
\texttt{end}
\texttt{end}

\texttt{\%newy=layer(1:length(othertime))- d_y;\% + mean(d_y);}
\texttt{\%layer = newy;}
\texttt{\% plotyy(othertime, layer(1:length(othertime)), othertime,}
\texttt{\% shear\_stress(1:length(othertime)))}
\texttt{\% plot(othertime,d_y, othertime,layer(1:length(othertime)),}
\texttt{\% othertime,newy);}
\texttt{for counter=1:length(d);}
\texttt{stickslip(counter,1)=rec(b(counter)); \%Rec at peak}
\texttt{stickslip(counter,2)=rec(d(counter)); \%Rec at min}
\texttt{stickslip(counter,3)=time(b(counter));}
\texttt{stickslip(counter,4)=time(d(counter));}
\texttt{stickslip(counter,5)=shear\_stress(b(counter));}
\texttt{stickslip(counter,6)=mean(shear\_stress(d(counter):d(counter)+20));}
\texttt{stickslip(counter,7)=friction(b(counter));}
\texttt{stickslip(counter,8)=mean(friction(d(counter):d(counter)+20));}
\texttt{stickslip(counter,9)=ec\_disp(b(counter));}
\texttt{stickslip(counter,10)=mean(ec\_disp(d(counter):d(counter)+20));}
\texttt{layerbefore(counter)=layer(b(counter));}
\texttt{layerafter(counter)=mean(layer(d(counter):d(counter)+20));}
\texttt{stickslip(counter,11)=layerbefore(counter);}
\texttt{stickslip(counter,12)=layerafter(counter);}
\texttt{end}

\texttt{\% Use the stored positions to make a master variable to be saved}
\texttt{\%stickslip(:,1)=rec(b); \%Rec at peak}
\texttt{\%stickslip(:,2)=rec(d); \%Rec at min}
\texttt{\%stickslip(:,3)=time(b); \%Time at peak}
\end{verbatim}
\end{verbatim}
%stickslip(:,4)=time(d); %Time at min
%stickslip(:,5)=shear_stress(b); %Stress at peak
%stickslip(:,6)=shear_stress(d); %Stress at min
%stickslip(:,7)=friction(b); %Friction at peak
%stickslip(:,8)=friction(d); %Friction at min
%stickslip(:,9)=ec_disp(b); %Displacement at peak
%stickslip(:,10)=ec_disp(d+3); %Displacement at min
dummytime=time(d+3); %Time for articual displacement

fname=strcat(name,'stickslips.txt'); %User defined experiment name above

if saver=='y'
    save (fname, 'stickslip', '-ASCII', '-tabs')
    Output=strvcat('Data saved to file:',fname)
else
    'No file created'
end

%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%
% Calculated parameters
% messes up the last stick slip so that dimensions will agree
%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%

if calc=='y'
    for n=1:length(stickslip)-1;
        totalslip(n)=stickslip(n+1,9)-stickslip(n,9);
        recurrence(n)=stickslip(n+1,3)-stickslip(n,3);
        stressdrop(n)=stickslip(n,5)-stickslip(n,6);
        seis(n)=stickslip(n,10)-stickslip(n,9);
        layerchange(n)=layerbefore(n)-layerafter(n);
    end
    preseis = totalslip-seis;
    seisenergy=1/2*stressdrop/1000/1000.*seis*1000*1000; %Joules/m2
    Es_v = seisenergy./layerafter(1:length(layerafter)-1);
    Es_a = seisenergy *area;
    totenergy=avestress(1:length(avestress)-1)/1000/1000.*totalslip...
    *1000*1000; %Joules
    W_v=totenergy./layerafter(1:length(layerafter)-1);
    W_a=totenergy*area;
    percentseis=seisenergy./totenergy*100; %No units
    layerenergy=normstress/1000/1000*layerchange*1000*1000;
    Wl_v = layerenergy./layerafter(1:length(layerafter)-1);
    Wl_a = layerenergy*area;
perct_tot = Es_v./(W_v + Wl_v)*100;

else
    'No calculations completed'
end

if plotter == 'y' & calc == 'y'
    figure(1)
    subplot(3,1,1)
    plot(stickslip(1:length(stickslip)-1,3),stressdrop, '.')
    xlim([0,100]); xlabel('Time at event (s)'); ylabel('Stress Drop (MPa)')
    title(['Experiment ', name, ' \sigma_n = ',num2str(normstress),'MPa', ...
          ' v = ', num2str(velocity), '\mu m/s'])
    subplot(3,1,3)
    plot(time,shear_stress)
    xlim([0,100]); xlabel('Time (s)'); ylabel('Shear Stress(MPa)')
    subplot(3,1,2)
    plot(stickslip(1:length(stickslip)-1,3),recurrence, '.')
    xlim([0,100]); ylim([0, 1]);
    xlabel('Time at event (s)'); ylabel('Recurrence(s)')
    plotname = strcat(name,'F1');
    hgsave(plotname)

    figure(2)
    plot(stickslip(:,3),stickslip(:,5),'rx',stickslip(:,4),... 
         :,stickslip(:,6),'go', time,shear_stress)
    legend('Peaks','Troughs', 4); xlim([0,100]);
    xlabel('Time (s)'); ylabel('Shear Stress(MPa)')
    title(['Experiment ', name, ' \sigma_n = ',num2str(normstress),'MPa', ... 
           ' v = ', num2str(velocity), '\mu m/s'])
    plotname = strcat(name,'F2');
    hgsave(plotname)

%stickslip(:,4)
figure(3)
plot(stickslip(:,3),stickslip(:,9),'rx',dummytime,... 
     :,dummytime(:,10),'go', time,ec_disp)
legend('Peaks','Troughs', 4); xlim([0,100]);
xlabel('Time (s)'); ylabel('Elastic Corrected Displacement (\mu m)')
```matlab
figure(4)
subplot(4,1,1)
plot(stickslip(1:length(stickslip)-1,3),seis,'.')
xlim([0,100]); ylim([-10 150]);
xlabel('Time at event (s)'); ylabel('Seismic Slip ($\mu m$)');
title(['Experiment ', name,' \sigma_n = ',num2str(normstress),'MPa','
     v = ', num2str(velocity), '{\mu m/s}'])
subplot(4,1,2)
plot(stickslip(1:length(stickslip)-1,3),totalslip,'.')
hold on
stem(stickslip(1:length(stickslip)-1,3),preseis,'r.')
xlim([0,100]); ylim([-10 150]); legend ('Total','Preseismic')
xlabel('Time at event (s)'); ylabel('Total Slip ($\mu m$)');
hold off
subplot(4,1,3)
plot(stickslip(1:length(stickslip)-1,3),layerchange,'.')
xlim([0,100]); ylim([0 500]);
xlabel('Time at event (s)'); ylabel('Layer Work (J/m2)');
subplot(4,1,4)
plot(time,shear_stress)
xlim([0,100]); xlabel('Time (s)'); ylabel('Shear Stress(MPa)');
plotname = strcat(name,'F4');
hgsave(plotname)

figure(5)
subplot(4,1,1)
plot(stickslip(1:length(stickslip)-1,3),seisenergy,'.')
xlim([0,100]);
xlabel('Time at event (s)'); ylabel('Seismic Energy, E_s (J/m2)');
title(['Experiment ', name,' \sigma_n = ',num2str(normstress),'MPa','
     v = ', num2str(velocity), '{\mu m/s}'])
subplot(4,1,2)
plot(stickslip(1:length(stickslip)-1,3),totenergy,'.')
xlim([0,100]);
xlabel('Time at event (s)'); ylabel('Total Work, W_{tot} (J/m2)');
subplot(4,1,3)
plot(stickslip(1:length(stickslip)-1,3),layerenergy,'.')
xlim([0,100]); ylim([0 500]);
xlabel('Time at event (s)'); ylabel('Layer Work (J/m2)');
subplot(4,1,4)
plot(stickslip(1:length(stickslip)-1,3),percentseis,'.')
```

198
xlim([0,100]); ylim([0,10]);
xlabel('Time at event (s)'); ylabel('E_s/W_{tot}*100')
plotname = strcat(name,'F5');
hgsave(plotname)

figure(6)
subplot(4,1,1)
plot(stickslip(1:length(stickslip)-1,3),Es_v, '.')
xlim([0,100]);
xlabel('Time at event (s)'); ylabel({'Seismic Energy, E_s (J/m^3)'}

subplot(4,1,2)
plot(stickslip(1:length(stickslip)-1,3),W_v, '.')

subplot(4,1,3)
plot(stickslip(1:length(stickslip)-1,3),Wl_v, '.')

subplot(4,1,4)
plot(stickslip(1:length(stickslip)-1,3),percentseis, '.')

figure(7)
subplot(4,1,1)
plot(stickslip(1:length(stickslip)-1,3),Es_a, '.')

subplot(4,1,2)
plot(stickslip(1:length(stickslip)-1,3),W_a, '.')

subplot(4,1,3)
plot(stickslip(1:length(stickslip)-1,3),Wl_a, '.')

subplot(4,1,4)
plot(stickslip(1:length(stickslip)-1,3),percentseis, '.')
plotname = strcat(name,'F7');
hgsave(plotname)

figure(8)
 subplot(3,1,1)
 plot(stickslip(1:length(stickslip)-1,3),Es_v,'.')
 xlim([0,100]); ylim([0 0.1]);
 xlabel('Time at event (s)'); ylabel({'Seismic Energy, E_s (J/m3)'})
 title([['Experiment ', name,' \sigma_n = ',num2str(normstress),'MPa','... 
 ' v = ', num2str(velocity), '{\mu}m/s'])
 subplot(3,1,2)
 plot(stickslip(1:length(stickslip)-1,3),W_v + Wl_v,'.')
 xlim([0,100]);
 xlabel('Time at event (s)'); ylabel('W_{tot} + W_{comp} (J/m3)')
 subplot(3,1,3)
 plot(stickslip(1:length(stickslip)-1,3),perct_tot,'.')
 xlim([0,100]); ylim([0,10]);
 xlabel('Time at event (s)'); ylabel('E_s/(W_{tot} + W_{comp})*100')
 plotname = strcat(name,'F8');
hgsave(plotname)

figure(9)
 plot(othertime,layer(1:length(othertime)), stickslip(:,3), ...
   layerbefore, 'o', stickslip(:,4), layerafter, 'x')
 xlabel('Time (s)'); ylabel('Layer Thickness (microns)');
 title([['Experiment ', name,' \sigma_n = ',num2str(normstress),'MPa','... 
 ' v = ', num2str(velocity), '{\mu}m/s'])
 plotname = strcat(name,'F9');
hgsave(plotname)

else
   'No plotting'
end
Figure 1. Stress drop, recurrence interval and shear stress against time.

Figure 2. Stress against time for the experiment with the picks that the code takes as the peak and trough.
Experiment pl 553  \( \sigma_0 = 40 \text{MPa} \)  \( V = 200 \text{mm/s} \)

- Total Work, \( W_{\text{tot}} \) (Seismic Energy / \( E_\text{f} \))
- Linear Work, \( W_\text{lin} \)
- \( W_\text{lin} + W_\text{surf} \) (kBm)
- \( E_\text{f} (W_{\text{tot}} + W_\text{surf})/100 \)
2. Creep rate calculator

clear all
close all

%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%
% Written by Andy Rathbun, Jun 24 2005
% Line Fitting Program to Read in and Calculate Strain Rate and
% Stress Exponent
%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%

p758=dlmread('/barre/s0/data/p758/p758.txt','
);
shear_stress=p758(:,3); %MPa
time=p758(:,7); % Sec
strain=p758(:,5);
name='p758';
clear p758

%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%

%inputs: Row number of the start and end of each disp

tau=[1502 2856; 2858 4211; 4214 5642; 5644 6996; 6998 8457; 8459 9811; ... 9813 10053];
stress=[0.435 0.462 0.488 0.515 0.541 0.567 0.593];
s_tau=[1502 2856; 2858 4211; 4214 5642; 5644 6996; 6998 8457; 8459 9811];
%Stable stresses that
%shold be fit
s_stress=[0.435 0.462 0.488 0.515 0.541 0.567];
strength=0.61658; %Shear stress at the end of the run in (MPa)
normstress=1.001; %MPa
%stress=stress/strength; %Normalize shear stress
number=length(s_stress); %Number of stress intervals, subtract 1 to keep out
%the tertiary unstable stress step
samplinginterval=10; %Samples/sec
decimation=20; %Decimation times
samp_sec=samplinginterval/decimation; %New samples per sec
min=2; %Input the number of minutes for your fit
min=min*60; %Convert to sec
points=min*samp_sec;
t1=20*60*samp_sec; t2=40*60*samp_sec; %Times for each fit to start In this
%case it is
%20 min, 40 min and the min interval before the end of timing for each stress

%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%
%Fitting section

i=1;
p1=[2, number];
p2=[2, number];
p3=[2, number];
A=[2, number];
B=[2, number];
sigmaY=[2, number];
sigmaA=[2, number];
sigmaB=[2, number];
for k=1:number
    for counter=(s_t(k,1)):1:(s_t(k,1)+points)
        j1=counter+t1;
        x(i) = time(j1);
        y(i) = strain(j1);
        j2=counter+t2;
        x2(i) = time(j2);
        y2(i) = strain(j2);
        i=i+1;
    end
    i=1;
    format short
    q=polyfit(x,y,1);
    p1(1,k)=q(1,1);
    p1(2,k)=q(1,2);
    q2=polyfit(x2,y2,1);
    p2(1,k)=q2(1,1);
    p2(2,k)=q2(1,2);

% My calculation of slope (B), intercept (A) and error (sigma)
% Makes a 2 by number matrix for A,B,sigmaY,sigmaA,sigmaB
% Top row is 20 m, bottom is 40 m

N=length(x);
N2=length(x2);
sumX=sum(x);
sumX2=sum(x2);
sumY=sum(y);
sumY2=sum(y2);
sumXsq=sum(x.*x);
sumX2sq=sum(x2.*x2);
sumXY=sum(x.*y);
sumXY2=sum(x2.*y2);
\[ \text{delta} = N \cdot \text{sumXsq} - \text{sumX}^2; \]
\[ \text{delta2} = N_2 \cdot \text{sumX2sq} - \text{sumX2}^2; \]
\[ A(1,k) = \frac{(\text{sumXsq} \cdot \text{sumY} - \text{sumX} \cdot \text{sumXY})}{\text{delta}}; \]
\[ \% \text{intercept 20 m} \]
\[ B(1,k) = \frac{(N \cdot \text{sumXY} - \text{sumX} \cdot \text{sumY})}{\text{delta}}; \]
\[ \% \text{slope 20 m} \]
\[ A(2,k) = \frac{(\text{sumX2sq} \cdot \text{sumY2} - \text{sumX2} \cdot \text{sumXY2})}{\text{delta}}; \]
\[ \% \text{intercept 40 m} \]
\[ B(2,k) = \frac{(N_2 \cdot \text{sumXY2} - \text{sumX2} \cdot \text{sumY2})}{\text{delta2}}; \]
\[ \% \text{slope 40 m} \]
\[ \text{sigmaY}(1,k) = \sqrt{\frac{1}{(N-2)} \cdot \sum ((y - A(1,k) - B(1,k)x)^2)}; \]
\[ \% \text{Caclulated error in y comes from residuals} \]
\[ \text{sigmaA}(1,k) = \text{sigmaY}(1,k) \cdot \sqrt{\frac{\text{sumXsq}}{\text{delta}}}; \]
\[ \% \text{value of intercept} \]
\[ \text{sigmaB}(1,k) = \text{sigmaY}(1,k) \cdot \sqrt{\frac{N}{\text{delta}}}; \]
\[ \% \text{value of slope} \]
\[ \text{sigmaY}(2,k) = \sqrt{\frac{1}{(N_2-2)} \cdot \sum ((y2 - A(2,k) - B(2,k)x2)^2)}; \]
\[ \% \text{sigmaY}(2,k) = 0.0004; \]
\[ \% \text{My estimate of error in y to use instead} \]
\[ \text{sigmaA}(2,k) = \text{sigmaY}(2,k) \cdot \sqrt{\frac{\text{sumX2sq}}{\text{delta2}}}; \]
\[ \% \text{sigmaA}(2,k) = \text{sigmaY}(2,k) \cdot \sqrt{\frac{N_2}{\text{delta2}}}; \]
\[ \% \text{sigmaA}(2,k) = \text{sigmaY}(2,k) \cdot \sqrt{\frac{N_2}{\text{delta2}}}; \]
\[ \% \text{sigmaB}(2,k) = \text{sigmaY}(2,k) \cdot \sqrt{\frac{N_2}{\text{delta2}}}; \]

\[ f1 = \text{polyval}(q,x); \]
\[ \text{figure}(); \]
\[ \text{subplot}(2,1,1) \]
\[ \text{plot}(x, y \cdot 100, 'o', x, f1 \cdot 100) \]
\[ \text{xlim}([x(1,1), \text{max}(x)]) \]
\[ \text{title('20 min after step','num2str(min/60),... } \]
\[ \text{' min fit \tau = ',', num2str(s\_stress(k)), 'MPa'}) \]
\[ \text{ylabel('Percent Strain')} \]
\[ \text{plusminus} = \text{sigmaB}(1,k); \]

\[ \text{if } q(1,1)<0 \]
\[ \text{stringmatrix}(1,1:5) = 'Data'; \]
\[ \text{stringmatrix}(2,1:9) = \text{num2str}(q(1,1), '%2.2e'); \]
\[ \text{stringmatrix}(2,10:12) = 'pm'; \]
\[ \text{stringmatrix}(2,13:19) = \text{num2str}(\text{plusminus}, '%2.1e'); \]
\[ \text{legend(stringmatrix,-1)} \]
\[ \text{else} \]
\[ \text{stringmatrix}(1,1:5) = 'Data'; \]
\[ \text{stringmatrix}(2,1:8) = \text{num2str}(q(1,1), '%2.2e'); \]
\[ \text{stringmatrix}(2,9:11) = 'pm'; \]
\[ \text{stringmatrix}(2,12:18) = \text{num2str}(\text{plusminus}, '%2.1e'); \]
\[ \text{legend(stringmatrix,-1)} \]
\[ \text{end} \]
\[ \text{clear stringmatrix} \]

\[ f2 = \text{polyval}(q2,x2); \]
\[ \text{subplot}(2,1,2) \]
\[ \text{plot}(x2, y2 \cdot 100, 'o', x2, f2 \cdot 100) \]
\[ \text{xlim}([x2(1,1), \text{max}(x2)]) \]

206
title(['40 min after step ',num2str(min/60),...
     ' min fit \tau = ', num2str(s_stress(k)), ' MPa'])
ylabel('Percent Strain')
plusminus=sigmaB(2,k);
if q2(1,1)<0
    stringmatrix(1,1:5)='Data '; 
    stringmatrix(2,1:9)=num2str(q2(1,1),'%2.2e');
    stringmatrix(2,10:12)='pm';
    stringmatrix(2,13:19)=num2str(plusminus,'%2.1e');
    legend(stringmatrix,-1)
else 
    stringmatrix(1,1:5)='Data '; 
    stringmatrix(2,1:8)=num2str(q2(1,1),'%2.2e');
    stringmatrix(2,9:11)='pm';
    stringmatrix(2,12:18)=num2str(plusminus,'%2.1e');
    legend(stringmatrix,-1)
end

ii=1;
clear xax 
clear yax 
for ploter=s_tau(k,1):1:s_tau(k,2)
    xax(ii)=time(ploter);
    yax(ii)=strain(ploter);
    ii=ii+1;
end
figure()
plot(x,f1*100,'xg',x2,f2*100,'xk',xax,yax*100)
title(['Fits and strain \tau = ', num2str(s_stress(k)),...
     ' MPa']), ylabel('Percent Strain')
xlabel('Time (s)')
    legend('20 min', '40 min', 'Entire Stress Step',4)
xlim([xax(1,1), max(xax)])
end

regres1=polyfit(log10(s_stress),log10(p1(1,:)),1);
% Fits the data points of stress
regres2=polyfit(log10(s_stress),log10(p2(1,:)),1);
and strain rate to find n and b

```matlab
fit1=10.^polyval((regres1),log10(s_stress));
fit2=10.^polyval((regres2),log10(s_stress));
figure()
loglog(s_stress,p1(1,:),'bx',s_stress,p2(1,:),'kd', s_stress,fit1,'b',...
s_stress, fit2, 'k')
reg1=num2str(regres1(1,1),'
reg2=num2str(regres2(1,1),'
xlabel('Shear Stress (MPa)'), ylabel('Strain Rate (1/s)')
title('p758 Saturated NO run in')
stringmatrix1(1,1:10)= '20 min n= ';
stringmatrix1(1,11:10+length(reg1))= reg1;
stringmatrix2(1,1:10)= '40 min n= ';
stringmatrix2(1,11:10+length(reg2))= reg2;
legend(stringmatrix1, stringmatrix2, 2)
hold on
errorbar(s_stress,B(1,:),sigmaB(1,:),'bx')
hold on
errorbar(s_stress,B(2,:),sigmaB(2,:),'ko')
hold off

[i]=strainplots(stress,tau,time,strain,name);
```

```matlab
function [i]=strainplots(stress,tau,time,strain,name)
    colors='rgbcmyk-+d';
    figure
    number=length(stress);
    for count=1:number
        i=1;
        clear x
        clear y
        for counter=(tau(count,1)):1:(tau(count,2))
            x(i) = time(counter);
            y(i) = strain(counter);
            i=i+1;
        end
        x=x-x(1);%Nomralize to have 0 as point of stress bump by subtracting
        y=y-y(1);
        plot(x,y,colors(count))
        hold on
        strs=num2str(stress(count),'%2.3f');
        stringmatr(count,1:length(strs))=strs
        stringmatr(count,(length(strs)+1):(length(strs)+4))=' MPa';
    end
```
end

title(name)
xlabel('Time (s)'), ylabel('Strain'), legend(stringmatr,2)
hold off
Andrew Paul Rathbun

Education
2010   Ph.D. The Pennsylvania State University, Chris Marone, Advisor
Dissertation:
2006   M.S., The Pennsylvania State University, Chris Marone, Advisor
Thesis: Laboratory Study of Till Deformation: Comparison of Creep and Strength Characteristics
2003   B.S. with Distinction in the Geological Sciences, The Ohio State University, Hallan Noltimier, Advisor

Professional Experience
2008   Research Scientist at ExxonMobil Upstream Research
2007   Teaching assistantship, Physical Processes in Geology, Penn State
2004-2005  Teaching assistantship, Oceanography, Penn State
2001-2004  Teaching assistant, Math 151, 152, 153, Ohio State
2003   Teaching assistant, Physical Geology, Ohio State
2002-2003  NSF REU for the McMurdo Dry Valleys, Antarctic LTER

Publications

Awards
2009 Euro Conference of Rock Deformation, Outstanding Young Scientist
2009 Penn State Geosciences Graduate Student Colloquium 1st Place Poster
2009 Shell Geosciences Energy Research Facilities Award
2007 ConocoPhilips Field Research Grant
2007 Penn State Geosciences Student Colloquium 2nd Place Pre-Comps PhD Talk
2005 Penn State Geosciences Student Colloquium 2nd Place Master’s Talk
2003 Sigma-Xi Grants-In-Aid of Research
2003 Ohio State Geology Department, Undergraduate Book Award
2003 Ohio State Honors College, Undergraduate Research Scholarship