Rock Strength of Caprock Seal Lithologies: Evidence for Past Seal Failure, Migration of Fluids and the Analysis of the Reservoir Seal Interface in Outcrop and the Subsurface

Elizabeth Sandra Petrie
Utah State University

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ROCK STRENGTH OF CAPROCK SEAL LITHOLOGIES: EVIDENCE FOR PAST SEAL FAILURE, MIGRATION OF FLUIDS AND THE ANALYSIS OF THE RESERVOIR SEAL INTERFACE IN OUTCROP AND THE SUBSURFACE

by

Elizabeth Sandra Petrie

A dissertation submitted in partial fulfillment of the requirements for the degree of

DOCTOR OF PHILOSOPHY

in

Geology

Approved:

Dr. James P. Evans
Major Professor

Dr. Peter Mozley
Committee Member

Dr. Susanne Janecke
Committee Member

Dr. Don Best
Committee Member

Dr. Joel Pederson
Committee Member

Dr. Mark R. McLellan
Vice President for Research and Dean of the School of Graduate Studies

UTAH STATE UNIVERSITY
Logan, Utah

2014
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ABSTRACT

Rock Strength of Caprock Seal Lithologies: Evidence for Past Seal Failure, Migration of Fluids and the Analysis of the Reservoir Seal Interface in Outcrop and the Subsurface

by

Elizabeth S. Petrie, Doctor of Philosophy

Utah State University, 2014

Major Professor: Dr. James P. Evans
Department: Geology

This research characterizes the nature of fractures in Paleozoic and Mesozoic caprock seal analogs exposed in central and south-eastern Utah. The results of this research show evidence for fluid flow and mineralization in the subsurface as well as reactivation of fractures suggesting that the fractures act as a loci for fluid flow through time. The heterolithic nature of the caprock seals and meso-scale (cm to m) variability in fracture distributions and morphology highlight the strong link between the variation in material properties and the response to changing stress conditions. The variable connectivity of fractures and the changes in fracture density at the meso-scale plays a critical role in subsurface fluid flow.

The presence or formation of new fractures can result in seal bypass systems, which can cause failure of hydrocarbon traps, CO₂ geosequestration sites, waste and
subsurface fluid repositories. An integrated approach of field, borehole geophysical, burial and stress history modeling, rock strength testing, and numerical modeling are used to understand the effects changing material properties, rock strength, and stress history have on sealing capacity.

Simplified stress history models derived from burial history curves are combined with laboratory derived rock properties to understand the importance variations in rock properties and differential and effective mean stress have on the mechanical failure of fine-grained clastic sedimentary rocks. Burial history and rock strength data show that in units that experience similar burial depths and changing mechanical property exert a control on deformation type. Geomechanical models reveal changes in local strain magnitudes at locked mechanical interfaces, suggesting that elastic mismatch between layers results in differential strain distribution.

Characterization of fracture patterns, rock strength variability and the modeled changes in subsurface strain distribution is especially important for understanding the response of low-permeability rocks to changing stress in the subsurface, and is applicable to multiple geo-engineering scenarios such as exploitation of natural resources, waste disposal, and management of fluids in the subsurface. The analyses presented in this dissertation provide analog fracture data for fine-grained clastic rocks and a dataset for better understanding the importance of heterogeneity in low permeability rocks.

(208 pages)
PUBLIC ABSTRACT

Rock Strength of Caprock Seal Lithologies: Evidence for Past Seal Failure, Migration of Fluids and the Analysis of the Reservoir Seal Interface in Outcrop and the Subsurface

by

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Utah State University, 2014

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Department: Geology

Scientists have proposed that in order to avoid damaging climate change further accumulations of anthropogenic atmospheric carbon dioxide (CO$_2$) must be limited. One of several proposed techniques for reducing the amount CO$_2$ reaching the atmosphere is carbon capture and storage (CCS). This emerging technology stores CO$_2$ emissions captured from large point sources (i.e., power plants or industrial facilities), in deep geologic formations, including depleted hydrocarbon reservoirs and saline aquifers. For successful CCS design, implementation, and appropriate site selection and subsurface trapping mechanisms must be ensured over the 100’s to 1000’s of year timescale.
A key component in trapping and storage of fluids or gas in the subsurface is caprock seal integrity. This dissertation research focuses on how variations in material properties, rock strength, and stress history of caprock seals affect sealing capacity. Compositional changes in rocks often concentrate near depositional interfaces and discontinuities, such as fractures and faults. These pre-existing structures can act as zones of weakness, which can be reactivated due to the injection and storage of fluids into the subsurface, and can result in increased permeability within the very fine-grained rocks that make up the sealing unit to subsurface fluid or gas. The results of this research show the important role variation in mechanical properties due to lithologic changes have on rock deformation and fracture formation and distribution.

Project costs were largely funded by a DOE research grant # DE-FC26-0xNT4 FE0001786 and DE-SC0004991 to Dr. J. P. Evans. Project costs were associated with travel to and from field sites with field assistants, sample preparation, geochemical and petrographic analysis of samples, as well as teaching assistantships, research assistantships, and a GDL Fellowship appointed to Elizabeth S. Petrie during the course of this research.
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A huge thank you to my committee members Peter Mozley, Joel Pederson, Susanne Janecke, and Don Best, for their patience, attention to detail, insightful advice, and inquisitive commentary throughout my time working on this dissertation.

I have been privileged to collaborate on several projects over the past 4 years and have learned an amazing amount from other scientists – Alvar Braathen, Niko Kampman, Stephen Bauer, and Joe Bishop are among a few who have inspired me to work harder and learn more. Dedicated field assistants and fellow graduate students including Rebecca Wood, Ryan Sontag, Kelly Bradbury, Corey Barton, Santiago Flores, Dave Richey, Mitchell Prante, Dawn Hayes, Katie Potter, Robin Nagy, and Natalie Burszytn whose commitment to geology and life outside geology has left its mark.

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Finally a huge amount of gratitude is owed to my family. Robert, who embraces the law of unintended consequences and helped turn it into Chapter 3, can find shade on the hottest of field days, and without whose companionship I would be lost. Last but not least Jemma and Maya, your laughter, sense of adventure, and ability to find joy everyday makes me smile.

Thank You.

Elizabeth Sandra Petrie
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CHAPTER 1

Introduction

Various geo-engineering applications require the injection of fluid into the subsurface for storage including: disposal of nuclear and chemical waste, wastewater from oil and gas operations, and CO\(_2\) captured from large point sources. Deep isolation and containment of these pressurized and/or buoyant fluids requires the presence of effective structural and/or stratigraphic traps and the presence of an effective caprock seal. This research evaluates the geo-mechanical response of changing stress conditions due to increased subsurface fluid pressure on the various rock types that comprise caprock seals and the reservoir seal interface.

In low-permeability clastic caprocks, fractures can control the location and volume of fluid flow in subsurface traps (Nelson, 2001; Cartwright et al., 2007). Leakage through open fractures or other discontinuities can occur over short geologic timescales (Nelson, 1999; Cartwright et al., 2007), and dominate fluid flow behavior in these rock types (National Research Council, 1996). This research evaluates the distribution of small-scale structures within caprock seal analogs that show evidence for past fluid migration. Fracture formation, propagation, and distribution patterns are controlled by lithologic anisotropy and structural history (Woodward and Rutherford, 1989; Nelson, 2001; Laubach et al., 2009). Structural history critical to understanding fracture propagation and distribution include:
stress history, the burial-deformation-uplift path, pre-existing fabrics/anisotropy, and structural position. Lithologic controls include mechanical properties associated with rock type, stratigraphic layering, porosity, and fluid content (Rijken and Cooke, 2001; Larsen et al., 2010).

Funding for this research was primarily provided for site screening and risk assessment of carbon capture and sequestration (CCS). CCS is a proposed mitigation technique for reducing future accumulations of anthropogenic CO$_2$ in the atmosphere by injecting CO$_2$ captured at large point sources such as coal fired power plants (Pacala and Socolow, 2004; Metz et al., 2005). Although the time scale for successful sequestration of CO$_2$ is on the order of 100’s to 1,000’s of years the most important mechanism for trapping of CO$_2$ on the immediate time scales (years to 10’s of years), when active injection is taking place. On these shorter times scales structural and/or stratigraphic traps are the dominant trapping mechanisms (Metz et al., 2005) (Fig. 1-1). Due to the compressibility of CO$_2$, storage would occur within geologic reservoirs deeper than 800 m, at temperatures greater than 31.1°C, with pH regulation based on formation lithologies and fluid chemistries at depth (Metz et al., 2005). At these conditions, super critical CO$_2$ is less dense than formation water and buoyant migration will occur away from the injection wellbore altering the fluid pressure at the reservoir seal interface. This increase in fluid pressure could result in mechanical failure of the overlying seal via creation of new hydraulic fractures or by reactivation of existing fractures or faults.
In addition to CCS, this research is applicable to production of hydrocarbons from unconventional reservoirs, such as tight sands and shale. Production from unconventional reservoirs requires the units be hydraulically fractured creating permeability pathways. Technically recoverable unconventional resources exhibit substantial increases in global distribution. Between 2011 and 2013 the EIA estimates a global increase of 9 % in shale gas and 110% shale / tight oil, with the United States ranking second in shale oil resources and fourth in shale gas resources (EIA, 2013).
Research Objectives

This dissertation evaluates sub-seismic seal by-pass systems (<10 m) to quantify the occurrence and morphology of fractures in caprock lithologies and across interfaces to understand how fractures interact with lithologic interfaces and how seal-by-pass systems are established. This research expands the scale of investigation from outcrop analogs to the subsurface through correlation of outcrop observations with elastic moduli derived from borehole wire-line data and finite element modeling strain distributions at subsurface conditions. This is an integrative project that uses geologic analogs to understand the importance of heterogeneity to mechanical rock failure. I combine outcrop datasets that evaluation of the nature, occurrence, and distribution of fluid flow pathways in the various sealing lithologies, with subsurface geophysical well bore data and laboratory derived rock strength data.

The specific research questions addressed here are:

1) How do lithologic changes within caprock seals and across the reservoir seal interface affect the distribution and morphology of fractures?

2) Do observed fractures show evidence of acting as leakage pathways?

3) How do changes in elastic moduli affect the strain distribution (locally) across stratigraphic boundaries and is this reflected in the distribution of fractures observed in outcrop?
4) What role do grain size and porosity play in the types of structures found in the damage zone?

5) How do changes in rock type affect the fluid pressure required for failure?

Study Localities and Research Approach

The geologic analogs used in this study are four caprock seals exposed in central and southeastern Utah in the central Colorado Plateau (Fig. 1-2). Each locality shows evidence for subsurface fluid flow across caprock seals and three of the localities contain exposures of the reservoir seal interface. Each analog has experienced a tectonic history that includes variable degrees of Sevier and Laramide deformation, normal faulting associated with halokinesis and salt dissolution, and then late Cenozoic exhumation associated with uplift of the Colorado Plateau and development of the Colorado River system.

All of the outcrop analogs evaluated show evidence of fluid flow along structural discontinuities and provide outcrop examples of micro- and meso-scale structures that represent the mechanical response to varied tectonic stress, diagenetic histories, and varied stratigraphic boundaries (interfaces). The outcrops examined include the Jurassic Carmel Formation and underlying Navajo Sandstone reservoir (Fig. 1-2A), exposed along I-70 in the San Rafael Swell, the Tununk Member of the Cretaceous Mancos Shale (Fig. 1-2B), where it is exposed near Crystal Geyser, the Earthy Member of the Jurassic Entrada Formation (Fig. 1-2C),
which contains intraformational reservoir seal pairs, where it is exposed at the Salt Wash Graben, and the Permian Organ Rock Formation and underlying Cedar Mesa Sandstone reservoir (Fig. 1-2D), exposed in Glen Canyon National Recreation Area.

This research combines outcrop surveys of discontinuity distribution, associated mineralization and stratigraphic changes in rock type to identify the relationships between composition, structural diagenesis/paragenesis and loading history. Field data documenting the occurrence of diagenetic alteration, fracture distribution, morphology of fractures, rock strength, and permeability surveys were gathered. Field data are combined with results from laboratory-based measurements of rock yield properties, which provide tensile strength data for estimation of the failure envelopes associated with each rock type. Elastic moduli derived from geophysical wireline data are used in combination with field observations of mechanical stratigraphy to build finite element models. The FEM are used to evaluate strain distribution across geo-mechanical boundaries. The model results are compared to outcrop observations of fracture distribution, with fractures being the manifestation of strain in outcrop.
Figure 1-2. Study location map, state of Utah geologic basemap from Hintze, et al., 2000. A) Jurassic Carmel Formation, B) Cretaceous Tununk Member of the Mancos Shale, C) Earthy Member of the Entrada Sandstone, D) Organ Rock Formation.
Dissertation Organization

Each chapter of this dissertation addresses a separate research component linked by addressing the theme of mechanical failure of caprock seals. They are presented in the sequence by which the research progressed.

Chapter 2 provides an overview of the mechanical stratigraphy as determined for the Carmel Formation and provides a method for deriving mechanical properties, specifically Young’s Modulus and Poison’s Ratio from borehole wireline well logs. This chapter focuses on correlations between outcrop observations of fracture distribution and morphology to subsurface derived elastic moduli with a view toward modeling the effects of changing elastic moduli on strain distributions. Chapter 2 is published as Petrie et al., 2012.

Chapter 3 addresses the effects grain size and porosity have on the formation of preferred permeability pathways and re-activation of structures within fault damage zones and provides evidence for flow within deformation band type faults in sandstones. Chapter 3 has been accepted for publication in the Journal of Structural Geology as Petrie et al.

Chapter 4 summarizes the field observations of the four failed seal analogs and uses results from tensile strength tests to compare the pore-fluid factor required to induce mechanical failure and failure type. Chapter 4 is in review in the American Association of Petroleum Geologists Bulletin.
Chapter 5 evaluates the elastic moduli derived in chapter II and uses finite element analysis to model the influence changes in elastic moduli have on strain distributions at interfaces defined by bedding. Chapter 5 has been presented as Petrie and Evans, 2013 at the Gussow Conference and the Unconventional Resources Technology Conference and is being prepared for submission to a special issue of the Bulletin of the Canadian Society of Petroleum Geologists.

In addition to the manuscripts presented in this dissertation, a Utah Geological Association guidebook article (Petrie et al., 2013) has been published. This article summarizes the research results carried out by the co-authors in the San Rafael Swell. Derivative products of this dissertation research have also been presented at several meetings see appendix figure A-1.

References Cited


CHAPTER 2

Predicting Rock Strength Variability at Stratigraphic Interfaces In Caprock Lithologies at Depth: Correlation Between Outcrop and Subsurface

Abstract

Open faults and fractures act as a major control of fluid flow in the subsurface, especially in fine-grained, low-permeability lithologies. These discontinuities often form part of seal bypass systems, which can lead to the failure of hydrocarbon traps, CO₂ geosequestration sites, waste and injected fluid repositories. We evaluate mesoscale variability in fracture density and morphology and variability in elastic moduli in the Jurassic Carmel Formation, a proposed seal to the underlying Navajo Sandstone for CO₂ geosequestration. By combining mechano-stratigraphic outcrop observations with elastic moduli derived from wireline log data, we characterize the variability in fracture pattern and morphology with the observed variability in rock strength within this heterolithic top seal.

Outcrop inventories of discontinuities show fracture densities decrease as bed thickness increases and fracture propagation morphology across lithologic interfaces vary with changing interface type. Dynamic elastic moduli, calculated from wireline log data, show that Young’s modulus varies by up to 40 GPa across depositional interfaces, and by an average of 3 GPa across the reservoir/seal
interface. We expect that the mesoscale changes in rock strength will affect the distributions of localized stress and thereby influence fracture propagation and fluid flow behavior within the seal. These data provide a means to closely tie outcrop observations to those derived from subsurface data and estimates of subsurface rock strength. The characterization of rock strength variability is especially important for modeling the response of caprocks to changing stress conditions associated with increased fluid pressures and will allow for better site screening and subsurface fluid management.

Introduction

For geosequestration of CO₂ to be viable, the formation-water saturated caprock must prevent migration of CO₂ from the injection reservoir into the shallow fresh-water aquifers, adjacent mineral deposits, fluid reservoirs or the biosphere. Caprocks or seals are fine-grained, low capillary-entry pressure units that serve as aquitards and prevent the upward migration of fluids (Gluyas and Swarbrick, 2003). Seal integrity is defined as the lack of compromising alterations, such as increased permeability due to lithologic changes or discontinuities within a seal that may affect the ability of a caprock to prevent fluid migration over geologic time scales (Bachu, 2008). Diffusion, capillary invasion and seal by-pass through connected discontinuities, such as fractures and faults, can result in seal failure. For CO₂ sequestration efforts, the rate of CO₂ infiltration into the extremely low permeability
caprock is expected to occur on geologic time scales and is therefore a non-issue for geosequestration schemes (Bennion and Bachu, 2008). The presence of structural discontinuities, such as fractures, a surface across which there is no cohesion, which in this paper includes joints, small displacement faults, and mineralized veins, within caprocks can result in seal bypass over the short time scales of years to 100's of years, prior to the time for which longer-term storage mechanisms such as dissolution and mineral trapping becoming dominant (IPCC 2005). Studying the occurrence of, and changes in, fracture patterns from outcrops or core samples, (mm to m), and scaling it up for use in modeling at the field scale, (m to km) for storage or production is difficult due to the lack of direct correlation of outcrop observations with subsurface data and the paucity of data between boreholes. Due to the amount of data needed in subsurface modeling scenarios carried out at the field scale, the mesoscale (cm- to m-scale) variability in rock properties is often neglected. However, variability at these scales may impact the initiation of fracture within the caprock.

The prediction of new fracture formation or reactivation of existing fractures in caprock lithologies at depth requires that we develop accurate geomechanical models that account for intra-seal lithologic variability and the response of caprocks to changing stress and fluid pressure conditions. Exposures of mudstone and siltstone sequences at several localities in central Utah show evidence of brittle-failure and fluid flow at depth (Heath et al., 2009; Dockrill and Shipton, 2010). These localities are typified by oxidation of the host rock along fracture boundaries
and commonly contain mineralized veins as well as surficial tufa and travertine deposits. Detailed evaluation of caprocks at proposed CO₂ sequestration sites should include tectonic, depositional, burial, and diagenetic histories (Li et al., 2006; Lu et al., 2011). The variability in these histories will affect the response of caprock lithologies to stress and should be implemented in modeling scenarios to understand the magnitude their effect has on fracture creation, propagation, or reactivation due to increased pore fluid pressures and the effect their presence has on fluid plume and pressure front migration.

Criteria for significant leakage and established methodologies for formal risk and performance assessments in CO₂ storage (Wilson and Gerard, 2007; Oldenburg 2008; NETL BPM, 2011) indicate that one of the greatest risks that may lead to caprock failure is the development and/or enhancement of caprock imperfections including fractures and connected fracture networks, faults, sedimentary facies changes, dissolution pipes, and the presence of intrusions (Almon and Dawson, 2004; Ingram et al., 2007; Pruess, 2008). These features are referred to as seal-bypass systems and are preferential flowpaths. Leakage scenarios for CO₂ portrayed by the Intergovernmental Panel on Climate Change (IPCC, 2005) assessment include large fault zones (Streit and Hillis, 2004; Shipton et al., 2005; Heath et al., 2009) and abandoned wellbores, but do not include networks of ubiquitous smaller fractures or sub-seismic faults.

In order to accurately model the mechanical (or poroelastic) behavior of a sequestration system, good constraints on the material properties, and their
variations, within the reservoir rock and its seals are needed. We evaluate the potential for fractures to exist in the subsurface, and determine the dynamic elastic moduli (Young’s modulus and Poisson’s ratio), from wireline log data to quantify the elastic strength / stiffness within different lithologies of a seal, and how these changes affect the strength of the entire seal. Ideally core would be available for evaluation, but no core over the Carmel Formation was available from either the United States Geological Survey Core Repository or the Utah Core Research Center, (Appendix Table B-1), within our study area. The interval of interest is cored in a borehole located in the Uinta Basin over 230 km (143 mi) away from the outcrop location, and due to the distance and difference in depositional and tectonic histories between the two areas, it was not evaluated in this study.

In this paper we establish a workflow for a more complete characterization of caprock integrity by correlating an exhumed caprock that shows evidence of past subsurface fluid flow with observed variability of dynamic elastic moduli calculated from wireline log data. Part of this workflow uses outcrop fracture interactions, cross-cutting relationships and fracture distributions at other outcrops, as well as petrography to establish the movement of fluids through open mode fractures in the subsurface. We use this workflow to quantify the mechanical properties of the Carmel Formation seal. Geomechanical modeling of at depth stress conditions using the results of the variability in dynamic elastic moduli will be carried out and presented in a later paper.
Study Area And Geologic Setting

The Middle to lower Upper Jurassic Carmel Formation provides an excellent opportunity to study a heterolithic low-permeability seal using both outcrop and subsurface data sets. The Carmel Formation is the proposed primary seal to the underlying Navajo and Page Sandstone sequestration targets and can be used as an analog for other heterolithic mixed carbonate siliciclastic seals (Allis et al., 2003; Esser et al., 2010; NETL, 2010). The Middle to lower Upper Jurassic Carmel Formation is well exposed in the San Rafael Swell, an asymmetric, east-vergent, doubly plunging anticline with a NNE-trending axis (Bump and Davis, 2003; Davis and Bump, 2009). Estimates on timing of formation of the San Rafael Swell range from 83.5 Ma to 58 Ma (Fouch et al., 1983; Lawton, 1986; Guiseppe and Heller, 1998; Shipton and Cowie, 2001). Structural lineaments trend ENE and WNW within the San Rafael Swell, with the hinge-line trending NNE (Figure 2-1). Modern maximum principal stress orientations vary slightly along gently dipping western limb of the San Rafael Swell from NNW (~350°) and shifts slightly to NNE (~8°) along to the Basin and Range/Colorado Plateau transition zone (Heidbach et al., 2008). At the study locality the Carmel Formation dips gently (9 ± 2°W) and unconformably overlies the Navajo Sandstone (Figures 2-1 and 2-2). The Carmel Formation is a mixed siliciclastic carbonate unit that was deposited in shallow marine to sabkha environments (Hintze, 1988; Blakey, 1994; Caputo, 2003).
The underlying Navajo Sandstone is a thick-bedded, well-sorted, very-fine to medium-grained permeable quartz sandstone that represents deposition in a large erg system (Hintze, 1988; Hansen, 2007). All four members of the Carmel Formation are exposed at this locality (Figure 2-2). These include: 1] the basal thin-bedded shale and sandy limestone of the Co-op Creek Member; 2] overlain by the medium to thick-bedded gypsiferous sandstone, mudstone and anhydrite layers of the Crystal Creek Member; 3] the siltstone and mudstones of the Paria River Member; and 4] the lower portion of the Windsor Member, interbedded micritic limestone and calcareous mudstone and siltstone (Hintze, 1988; Caputo, 2003; Sprinkel, 2010, personal communication) (Figures 2-2 and 2-3). The Co-op Creek and lower Crystal Creek Members make up the basal 9 m (29.5 ft) of the Carmel Formation and are considered in this study to be the primary seal to the underlying Navajo Sandstone. This portion of the Carmel Formation is dominated by thin- to medium-bedded bioclastic micritic limestone, which contains very fine to fine-grained quartz clasts and interbedded clay-rich, fissile to tabular siltstone and mudstone (Figure 2-2 and 2-3).

The Carmel Formation exhibits low permeability (from 1.61×10^-15 m² to 5.2×10^-14 m² / 0.002 Darcy to 0.053 Darcy), and in outcrop, contains near vertical mineralized veins, indicating that open mode fractures formed and were the loci of fluid flow at depth (Figure 2-3). Through-going mineralized fractures are observed throughout the Carmel Formation at this and other localities. These mineralized veins provide evidence for past seal failure, and suggests that the Carmel Formation
could fail under increased pore-fluid pressure, along existing discontinuities, resulting in the creation of a seal bypass system.

The inherent lithologic heterogeneity of this seal also imparts varied mechanical properties over small scales (10 cm to 1 m). This heterogeneity adds to the complexity of this unit and may require detailed modeling scenarios for accurate evaluation of its seal capacity and integrity. The calcite veins are in the resistant lithologies (limestone) while fractures in the finer grained lithologies are characterized by the presence of limonite mineralization calcite and/or gypsum mineralization.

We examined data from twenty offset boreholes (Figure 2-1), the nearest of which is 5 km to the NNW of the outcrop locality. In this wellbore the top of the Carmel Formation is interpreted to occur at 1640 m (5,381 ft) elevation above sea level. Wireline well-log data throughout Utah shows that the Carmel Formation thickens westward, ranging from 40-120 m (131-394 ft) thick. In general, the wireline log signatures used in this study show an eastward decrease in bed-thickness and loss of the upward fining character of beds.
Figure 2-1. Location of study area within Utah with tectonic elements map with borehole and outcrop locations examined in this study. Gray area is extent of Jurassic outcrop in the San Rafael Swell. Highlighted borehole locations are discussed further in text and Figure 2-8.
Figure 2-2. Carmel Formation and Navajo Sandstone outcrop at the I-70 exposure. Members of the Carmel Formation listed at right. Upper portion of the Windsor Member exposed on south side of I-70. A total of 38 m of Carmel Formation is exposed at this locality. The uppermost portion of the Windsor Member is a slope-forming shale and siltstone.
These wireline log observations are consistent with the interpreted changes in depositional environment from east to west that characterized Utah during the Middle Jurassic, with establishment of the Western Interior seaway resulting in a shift from widespread aeolian deposition in the east to near-shore/marine deposition in the west (Hintze, 1988; Blakey and Ranney, 2008). The lateral facies changes and resulting heterogeneity of the Carmel Formation are expected to have an impact on overall rock strength estimates made from wireline logs, as well as the fracture patterns observed in outcrop.

Methods

Data collected at the outcrop location includes fracture orientation, distribution, termination, length and mineralization as well as field-derived compressive strength and permeability data, modified from those used to calculate rock mass ratings (Bieniawski, 1989; Priest, 1993; Zhang, 2005). These field data are used in combination with the quantitative stratigraphic analysis of fracture density distributions (after Bertotti et al., 2007), from scan line fracture inventories (La Pointe and Hudson, 1985), and digital outcrop orthoimage analysis, to delineate mechano-stratigraphic units. The aim of combining criteria similar to those used in rock mass strength ratings with the stratigraphic analysis of fracture distribution is to identify areas within the heterolithic seal where changes in lithology(s) result in significant changes in deformation behavior. Changes in deformation behavior are
used here to identify individual mechano-stratigraphic units, which show a similarity in fracture distribution, permeability, rock strength and stratigraphic stacking patterns. Using this systematic data set we are able to split the outcrop into five quantitatively defined mechano-stratigraphic units (Figure 2-4) (Table 2-1), defined by their consistency in fracture distribution, bed thickness, lithologic stacking pattern, field-derived compressive strength, and permeability data.

Discontinuity data, including fault and fracture orientation, distribution, termination, length and mineralization data were obtained using scan lines, 2x2 meter window surveys and photogrammetry. Photogrammetry was done by compiling digital outcrop photo pairs as orthoimages using SiroVision®. SiroVision® is a software package that renders stereo pairs and allows for georeferenced interpretation of fracture orientation, length, fracture termination and fracture interactions. An N-type Schmidt hammer was used to collect estimates of outcrop compressive strength over the stratigraphic section; detailed methodology for these measurements is given in Selby (1993). Outcrop permeability data was collected through the exposed stratigraphic section using a TinyPerm II field permeameter. Both the outcrop derived permeability and compressive strength measurements were done on cleaned surfaces away from cracks or edges of outcrop to limit the effects of adjacent free surfaces; these data were used to characterized the outcrop for mechano-stratigraphic divisions and likely overestimate permeability (Fossen et al., 2011), and underestimate compressive strength (Selby, 1993).
Figure 2-3. Measured stratigraphic section with Carmel Formation member names. Detailed photos with sketch maps of fracture morphology across lithologic interfaces.
Figure 2-4. Stratigraphic column and associated mechano-stratigraphic field data compiled from outcrop scan lines and orthoimages. These data include bed thickness and fracture density as well as field-derived permeability and compressive strength.
Table 2-1. Mechano-stratigraphic data compiled from outcrop observations

<table>
<thead>
<tr>
<th>Mechano-stratigraphic unit</th>
<th>Median bed thickness (m/ft)</th>
<th>Mean fracture density ± std.dev. (fracture/m)</th>
<th>Median fracture spacing</th>
<th>Compressive Strength range</th>
<th>Permeability (m² / Darcy) range</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>mean</td>
</tr>
<tr>
<td>1</td>
<td>9.00/29.53</td>
<td>1.30 ± 0.99</td>
<td>0.8</td>
<td>21-34</td>
<td>0.4x10⁻¹³ - 1.97x10⁻¹¹ / 0.04 - 19.9</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>28</td>
</tr>
<tr>
<td>2</td>
<td>0.20/0.66</td>
<td>5.49 ± 2.55</td>
<td>0.5</td>
<td>21-50</td>
<td>1x10⁻¹⁴ - 5x10⁻¹⁴ / 0.01 - 0.05</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>39</td>
</tr>
<tr>
<td>3</td>
<td>0.15/0.49</td>
<td>6.16 ± 1.74</td>
<td>0.7</td>
<td>32-54</td>
<td>2x10⁻¹⁴ - 4.5x10⁻¹³ / 0.12 - 0.46</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>44</td>
</tr>
<tr>
<td>4</td>
<td>1.02/3.35</td>
<td>1.05 ± 0.28</td>
<td>1.5</td>
<td>14-42</td>
<td>4x10⁻¹⁴ - 2.7x10⁻¹³ / 0.04 - 0.27</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>29</td>
</tr>
<tr>
<td>5</td>
<td>0.59/1.94</td>
<td>2.45 ± 1.82</td>
<td>2.2</td>
<td>22-68</td>
<td>6x10⁻¹⁵ - 9x10⁻¹⁴ / 0.01 - 0.91</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>59</td>
</tr>
</tbody>
</table>

These field derived values should not be extrapolated to assumed subsurface conditions, but do provide outcrop data essential in delimiting variations in mechanical stratigraphy. Details of the Schmidt Hammer and TinyPerm II specifications and their use at the outcrop are provided in Appendix B-2.

Wireline well-log analysis was carried out using 2D/3D EarthPak in the SMTKingdom™ Suite. This analysis included the digitization of wireline logs from 20 exploration boreholes and production wells covering approximately 560 km² (348
mi²) (Figure 2-1 and Appendix Table B-1). The depth (elevation referenced to sea level) of the top Carmel Formation in these boreholes ranges from 105 to 5400 m (344.5 to 17,717 ft) (Appendix Table B-1). Gamma ray, sonic and bulk density (where available) logs were digitized for the upper 10 m of the Navajo Sandstone to the base Entrada Sandstone. Gamma ray (GR) logs serve as a proxy for lithology, with finer grained lithologies (e.g. shale) typically having a higher GR value due to the increased presence of naturally radioactive elements. Sonic logs (DT) are acoustic wireline logs that can be used to calculate sonic velocity in rocks. Most borehole sonic logs in this study were measured with a monopole source from which only a compressional velocity value can be derived directly. Density logs (RHOB), which provide a record of the bulk (fluid and matrix) density were digitized and used in the calculation of Young’s modulus (Figure 2-3 and Appendix B-3); in cases where bulk density logs were not available density was calculated from sonic log data (Appendix B-3 Equation 2).

The calculation of the elastic parameters, Poisson’s ratio and Young’s modulus requires $V_p$ and $V_s$ data that can be obtained directly from dipole sonic logs (Appendix B-3; Equations 3 and 4). Previous workers used global data sets of laboratory measurements, seismic data and wireline well-log data to established empirical relationships between $V_p$ and $V_s$ (Pickett, 1963; Castanga et al., 1985) (Table 2-2). These data show that specific lithologies and mineralogies have specific $V_p$-to-$V_s$ relationships. Using data from two control wells in the Drunkards Wash field, Utah D7 (API: 43-015-30338) and Utah D8 (API: 43-007-30431), which
contain dipole sonic data, we derive \( V_p \) and \( V_s \) directly from log data. Data from these wells are used to evaluate the validity of deriving \( V_s \) from \( V_p \) data alone (Table 2-2). The \( V_p \)-to-\( V_s \) relationships established in the two control is consistent with those previously established (Pickett, 1963; Castanga et al., 1985), and was then used to calculate \( V_s \) from the available \( V_p \) logs in all other boreholes (Table 2-2 and Appendices Table B-1 and B-3).

We use the digitized wireline well log data to calculate dynamic values for Young’s modulus and Poisson’s ratio. Dynamic values for Young’s modulus and Poisson’s ratio are derived from elastic-wave velocity and density, such as those introduced by an acoustic compressional and shear wave signal. Static values for Young’s modulus and Poisson’s ratio are those that are derived from laboratory deformation experiments. We estimate dynamic value of elastic moduli within the caprock and across the reservoir caprock interface by building on observations made by previous workers of specific empirical relationships between lithology or mineralogy to compressional velocity (\( V_p \)) and shear velocity (\( V_s \)) (Pickett, 1963; Castanga et al., 1985; McCann and Entwisle, 1992). McCann and Entwisle (1992) show dynamic elastic moduli can provide a reasonable estimate of engineering rock properties, with dynamic elastic moduli and static elastic moduli having correlation coefficients ranging between 0.75 and 0.9 (McCann and Entwisle, 1992).
Table 2-2. \(V_p\)-to-\(V_s\) trends established by this study and \(V_p\)-to-\(V_s\) relationships established by previous workers.

<table>
<thead>
<tr>
<th>Cross-Plot</th>
<th>Gamma Ray</th>
<th>(V_p/V_s)</th>
<th>Lithology</th>
<th>(V_p/V_s)</th>
<th>Mineralogy</th>
<th>(V_p/V_s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>GR&lt;50 Carmel Formation</td>
<td>1.9</td>
<td>Limestone</td>
<td>1.9</td>
<td>Clay</td>
<td>2.0</td>
</tr>
<tr>
<td>B</td>
<td>150&gt;GR&gt;50</td>
<td>1.8</td>
<td>Dolomite</td>
<td>1.8</td>
<td>Quartz</td>
<td>1.5</td>
</tr>
<tr>
<td>C</td>
<td>GR&lt;50 Navajo Sandstone</td>
<td>1.7</td>
<td>Clean Sand</td>
<td>1.7</td>
<td>Calcite</td>
<td>2.0</td>
</tr>
<tr>
<td></td>
<td>GR&gt;150</td>
<td>1.5</td>
<td>Limey Sands</td>
<td>1.6</td>
<td>Dolomite</td>
<td>1.75</td>
</tr>
</tbody>
</table>

Results

Outcrop Analysis

Based on the field data collected, we define 5 mechanical units over a vertical stratigraphic distance of 38 m (125 ft) in the upper Navajo Sandstone and the lower Carmel Formation. The lowermost mechanical unit, unit 1 (Figure 2-4), is the uppermost Navajo Sandstone. The Navajo Sandstone has a mean fracture density of two fractures per meter (fracture/m) and contains open joints and fault deformation bands. The overlying Carmel Formation is split into four mechano-stratigraphic units, labeled 2-5 (Figure 2-4). Unit 2 is composed of interbeds of thin-bedded argillaceous limestone, siltstone and shale; this unit coarsens upward and is more fossiliferous up-section. A 1.5 m (5 ft) thick shale bed marks the base of unit 3. This unit is finer grained; more thinly bedded and has a higher fracture density than the underlying unit 2 (Figure 2-4; Table 2-1). Mechano-stratigraphic unit 4 has the lowest fracture density observed at this outcrop; it is a thin- to medium-bedded gypsiferous mudstone and sandstone unit capped by a meter scale gypsum bed
(Figure 2-4, Unit 4; Table 2-1). The uppermost mechano-stratigraphic unit, (Unit 5), is a thinly bedded micritic limestone and lime siltstone, it lacks the interbedded nature of units 2 and 3 below (Figure 2-3, Unit 5; Table 2-1).

![Graph showing fracture density versus bed thickness](image)

Figure 2-5. Fracture density versus bed thickness shows the expected relationship of lower fracture density in thicker beds. Exponential best-fit line with a correlation coefficient (R) of 0.66.

The Carmel Formation follows the expected relationship of an overall decrease in fracture density with increase in bed thickness (Figure 3C and 5) (Bai et al, 2000; Bai and Pollard 2000; Gross, 1993). Fracture density decreases in beds thicker than 0.5 meters (Figure 2-5), where fracture density ranges between 0.4 and 6.6 fractures/m with a mean fracture density of 2 ± 1.8 fracture/m. When bed thickness is less than 0.5 m, both weak and resistant units show a wide variability in
fracture density ranging from 0.11 to 9.72 fracture/m, with a mean fracture density of 5.7 ± 2.3 fracture/m (Figure 2-5) (Table 2-1).

Field-derived rock strength and permeability estimates also vary stratigraphically. Compressive rock strength, from Schmidt Hammer rebound values, exhibits higher variability in the thinly bedded heterolithic portion of the lower Carmel Formation, and a decrease in compressive strength in mechano-stratigraphic unit 4 and an increase in compressive strength in unit 5 (Table 2-1). The Carmel Formation is a low-permeability unit exhibiting an overall decrease in permeability up-section. The permeability values of less than 7 x 10⁻¹⁵ m² (0.007 Darcy), are recorded throughout the micritic limestone beds of mechano-stratigraphic unit 5 whereas permeability ranges from 16 x 10⁻¹⁴ m² to 5.2 x 10⁻¹⁴ m², (0.002 to 0.52 Darcy) in units 2 through 4 (Figure 2-4) (Table 2-1).

Permeability within the Navajo Sandstone 1 m (3.28 ft) below the reservoir/seal interface is orders of magnitude higher than the overlying Carmel Formation, permeability measurements within the reservoir have a mean value of 8.7 x 10⁻¹² m², (8.8 Darcy), (Figure 2-4) (Table 2-1).

There are two local fracture strike orientations in this area (F1 and F2). The Navajo Sandstone and Carmel Formation share NNW (F1) and NNE (F2) orientations. The dominant fracture strike is NNW, and the majority of fractures within the Carmel Formation, including veins, and fault deformation bands in the Navajo Sandstone share this orientation (Figure 2-6). Within the Navajo Sandstone, the open joint set has a dominant NNE, F2, strike orientation (mean strike 335°),
whereas deformation bands and shear fractures have a mean strike orientation of 21°.

We interpret the NNE fracture set in both the Navajo Sandstone and overlying Carmel Formation to reflect the modern maximum principal stress orientation (Heidbach et al., 2008), and are likely related to Cenozoic tectonic uplift and exhumation. The mineralized fractures observed in the Carmel Formation and the fault deformation bands in the Navajo Sandstone likely represent paleo-stress directions remnant of the Laramide uplift and associated deformation along the western edge of the San Rafael Swell (Anderson and Barnhard, 1986; Davis and Bump, 2009). The similar orientation of discontinuities and deformation band trends in the Carmel Formation and Navajo Sandstone suggests a similar history of formation and deformation (Figure 2-6). In thin-section mechanical twins are observed within the calcite veins of the Carmel Formation; the presence of mechanical twins indicates that fracture opening, mineralization and subsequent deformation occurred at depth.

The shale horizons of the Carmel Formation exhibit a greater dispersion in fracture orientation data than the limestone and sandstone lithologies (Figure 2-6). This dispersion and its association with lithology indicate that each rock type responds differently to stress, and in a stratigraphic sequence with high lithologic variability highly variable fracture distributions are likely (Figure 2-6). The distribution of bulls-eye patterns shown in the shale data (Figure 2-6) suggest that its presence and the distribution of this lithology within the sealing facies affect
fracture patterns and propagation morphology, which will in turn affect fluid flow at depth.

Variability in fracture propagation morphology is also evident within individual beds and across stratigraphic interfaces (Figure 2-3). This variability may be due to changes in stress localization at lithologic boundaries (Larsen et al., 2010). These morphologies include continuous fractures, fracture termination, fracture step-over, fracture swarm behavior and spaced fractures (Figure 2-3).

The changes in fracture morphology across stratigraphic interfaces will affect fluid flow in the subsurface due to changes in connectivity of open fractures across stratigraphic boundaries.

**Wireline Log Analysis**

We observe $V_p$-to-$V_s$ relationships within our control wells over specific GR values that are comparable to those established by previous workers for specific lithologies and mineralogies (Pickett, 1963; Castanga et al., 1985, Ellis and Singer, 2007) (Figure 2-7) (Table 2-1). Evaluation of the Navajo Sandstone and Carmel Formation in two Drunkards Wash wells show three clear $V_p$-to-$V_s$ relationships that can be grouped by their GR value ranges, with GR serving as proxy for lithology. This suggests that $V_s$ can be derived from the $V_p$ data based on GR values (Figure 2-7). The $V_p$-to-$V_s$ relationships coded by their GR groups, from the control well Utah D7, exhibit similar groupings in the Utah D8 borehole (Figure 2-7). The GR values of
less than 50 API and a $V_p$-to-$V_s$ relationship of 1.9 are characteristic of the argillaceous limestone and limestone facies of the Carmel Formation. The GR values between 50 and 150 API have a $V_p$-to-$V_s$ relationship of 1.8 and are characteristic of the fine-grained facies in the Carmel Formation. The GR values less than 50 API, having a $V_p$-to-$V_s$ relationships of 1.6, are characteristic of the Navajo Sandstone (Figure 2-7) (Table 2-1).

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Figure 2-6. Fracture orientation data from scan-lines in the Carmel Formation and Navajo Sandstone. Top: A, B, and C stereograms are equal area lower hemisphere projections of poles to fracture planes. Kamb Contours have a contour interval of 2.0 sigma. Bottom: half-circle rose diagrams of the same data. A) All Carmel Formation data; B) Carmel Formation fracture orientation data from resistant argillaceous limestone beds; C) Carmel Formation fracture orientation data from shale beds; D) Joints and fault deformation band trends measured with scan lines in the underlying Navajo Sandstone, data taken from cross-sectional and bedding plane exposures.
Given the relationships observed in our control wells (Figure 2-7), the consistency of this relationship with those previously published (Pickett, 1963; Castanga et al. 1985, Ellis and Singer, 2007), and our ability to replicate $V_s$ data from $V_p$ log data (Figure 2-8A), we can derive an estimate of shear velocity from $V_p$ and in turn calculate dynamic values for Poisson’s ratio and Young’s modulus for the rocks from compressional velocity wireline log data alone. Operations used to derive $V_s$ from $V_p$ values and elastic moduli are given in Appendix B-3. The calculation of shear velocity from compressional velocity in the control wells at Drunkards Wash resulted in no greater than 3% difference in the Carmel Formation and a difference of less than 4% in the Navajo Sandstone between values derived directly from dipole sonic log measurements and those calculated using the observed $V_p$-to-$V_s$ and GR relationships (Figure 2-8A).

The linear relationships used in the calculations of shear velocity results in Poisson’s ratio displaying an average value over each GR interval (Figure 2-8B; Appendix B-3 Equation 3). Calculations of dynamic Young’s modulus, in the basal 10 m of the Carmel Formation in the Celsius Fed 8-1 drill hole shows variability from 5 to 34 GPa at the meter scale and across interfaces within the Carmel Formation (Figure 2-7, 2-8, and 2-10; Appendix B-3, Equations 3 and 4). The calculated values for dynamic Young’s modulus in all wells results in a median value of 27.9 GPa and a mean value of 27.7 GPa ± 6.6 GPa. Outcrop observations show that these boundaries have variable fracture connectivity, fracture density and rock strength across stratigraphic boundaries.
These observations suggest that variability in fracture distribution and propagation morphology is the result of changes in elastic moduli across stratigraphic interfaces due to changes in the local stress concentrations at/across stratigraphic interfaces (Figure 2-9).

The changes in dynamic Young’s modulus across the reservoir/seal interface, calculated as the difference in the average dynamic Young’s modulus over 0.61 m (2 ft) on either side of the contact, ranging from 1 to 12 GPa, is relatively low when compared to shifts of up to 40 GPa in dynamic Young’s modulus observed across intra-seal interfaces (Figure 2-10). The difference in dynamic Young’s modulus across intra-seal interfaces can range as low as 0.04 GPa and as high as 22 GPa, when averaged over lithologic units. When peak to peak shifts, (i.e minimum to maximum Young’s modulus values), across the lithologic units are evaluated larger ranges in dynamic Young’s modulus exist within each borehole and range from 0.07 to 40 GPa (Figure 2-10).
Figure 2-7. Vp-to-Vs cross-plot with gamma-ray relationship in well Utah D7, well location highlighted in Figure 1. Group A) GR < 50 API, Vp/Vs = 1.9 shows dominantly low GR values in Carmel Formation; Group B) 50<GR<150 API, Vp/Vs 1.8 shows fine grained facies, high GR values in the Carmel Formation; Group C) GR < 50 API, Vp/Vs = 1.6 shows low GR values in Navajo Sandstone.
Figure 2-8. Well log calculation results. Left: Utah D7 wireline log, Vs log is measured shear velocity from dipole sonic; Vs is calculated shear velocity Right: Comparison of elastic moduli derived from control wells Utah D7, Utah D8 and Celsius Fed 8-1 and closest borehole to outcrop locations (5 km southeast of borehole). Borehole locations highlighted on Figure 2-1. In wireline logs shown JrC is top Carmel Formation, JcS is an intra-Carmel Formation marker, Jn is top Navajo Sandstone.
Discussion

We show that the nominally low permeability Carmel Formation contains complex fracture patterns and fracture morphologies associated with its changing rock types (Figures 2-3 and 2-4). Previous research has explored the influence of different types of lithologic interfaces on fracture morphology, termination, step-over or propagation (Cooke and Underwood, 2001; Larsen et al., 2010). Cooke and Underwood (2001) used outcrop analysis and numerical modeling to show that, at a moderate-strength interface, one that is moderately cemented or bonded, fractures will terminate or produce a step-over fracture; very weakly bonded contacts cause termination and strong contacts result in propagation across a boundary. We observe step-over fractures with and without bifurcation as well as termination and continuation of fractures at bedding interfaces within this outcrop. Two distinct fractures sets were observed in this study, one associated with deformation and fluid flow in the subsurface, F1, and the other linked to exhumation, F2, (Figure 2-6). Mechanical twins have been observed in the calcite veins of F1, which establishes the presence of open fractures that have been mineralized at depth. Fracture density distribution data shows the expected relationship of higher fracture densities associated with thin beds, resistant lithologies and adjacent to small offset (cm-scale) faults.

Evaluation of elastic moduli derived from wireline log data shows significant variability across the reservoir/seal interface and the intra-seal boundaries of the
Carmel Formation. Differences in dynamic Young’s modulus across the reservoir/seal interface range from -1.3 to 12.4 GPa, in general the difference across the reservoir-seal interface is positive, suggesting that the Carmel Formation has higher fracture strength than the underlying Navajo Sandstone (Figure 2-10). There are some wells, however, which show very little to no change across this boundary or negative values, suggesting that the Carmel Formation is more, or equally, prone to fracture than the underlying Navajo Sandstone (Figure 2-10).

A larger range of variability in dynamic Young’s modulus was observed across intra-seal lithologic units. These variations are comparable to the variability observed in the fracture pattern distributions throughout in the outcrop (Figures 9 and 10). For example, the dynamic Young’s modulus values derived from wireline well log data show changes from 5 to 34 GPa, peak to peak, across the stratigraphic interfaces within the lowermost Carmel Formation. In outcrop, fracture density varies from 4 to 7 fracture/m over the same portion of the Carmel Formation (Figure 2-9). Using the combined data sets from outcrop and wireline well logs, we will be able to populate a mechanical model with elastic properties averaged over the meter scale while our outcrop observations and delineation of mechano-stratigraphic units allows us to populate a mechanical model with layer thickness and lithologic stacking patterns (Figure 2-9). In our future modeling efforts we will combine the dynamic elastic moduli data with the outcrop data to identify the changes in subsurface rock properties that result in changes to local stress concentrations and are the ultimate control on both subsurface fracture distribution and connectivity.
Conclusions

The excellent exposure of the Carmel Formation in central Utah and the availability of wireline logs from nearby exploration boreholes and production wells provides a unique opportunity to tie outcrop observations of stratigraphic stacking and fracture distributions with rock strength variability derived from wireline log data. The combination of field- and geophysical-based methods show meso-scale variability in fracture distribution, fracture propagation, morphology and elastic moduli throughout the Carmel Formation. Our geophysical log data show significant changes in dynamic elastic moduli throughout the heterolithic seal, and field observations show continuous mineralized fractures well above the reservoir/seal interface, suggesting fluid flow through open fractures well into the sealing unit.

The variability in dynamic Young’s modulus across the reservoir/seal interface suggests that mechanical modeling is needed to determine how great a change (decrease) in Young’s modulus is required to arrest fractures at the reservoir/seal interface and understand the fracturability of the seal with changing lithology across intra-seal interfaces. Variations in elastic moduli within the Carmel Formation are associated with the heterolithic character of this unit (Figure 2-9).

The effects of this lithologic variability should be modeled to understand the severity of its influence on fracture generation and propagation. The variability in rock strength across the reservoir/seal interface and intra-seal interfaces is expected to affect local stress distributions and fracture propagation, fluid plume
and pressure front migration, as well as the creation of seal bypass corridors. The effects of this lithologic variability should be modeled to understand the severity of its influence on fracture generation and propagation. Variability in rock strength across sedimentary interfaces is expected to impact fluid flow and pressure distributions in the subsurface during the injection and early storage phase of CO₂ geosequestration projects.

We demonstrate that field-based methods can be used to define mechanostratigraphic relationships as well as to understand fracture distribution and timing. These field-based observations can be combined with wireline log derived variations in elastic moduli to further define zones of changing rock deformation behavior for use in mechanical modeling of rock response to changing stress conditions in the subsurface. In mechanical modeling, standard seal integrity analysis often assumes a constant mean value for a sealing unit, our field observations and the variability observed in derived values of dynamic elastic moduli, (Figure 2-9), suggests that mechanical modeling scenarios should attempt to capture smaller scale variations.
Figure 2-9. Comparison of potential model layers created by combining outcrop and wireline log datasets. Two possible model layers are presented here: Model 1 is based on shifts observed in the GR and $E_d$ on the wireline log. These model layers represent the median Young’s modulus, YM, value in that shift, and an average across the mechanical units, respectively. The second potential layering model, Model 2, is based on median values of calculated $E_d$ over 0.6 m intervals, the standard source/receiver spacing distance on the sonic wireline tool.
Figure 2-10. Top: Differences in dynamic Young’s modulus across the reservoir/seal interface. Calculated as the difference in average dynamic Young’s modulus over 61 cm (2 ft) in the uppermost Navajo Formation and 61 cm (2 ft) in the lowermost Carmel Formation. Middle: Differences in dynamic Young’s modulus across the intra-seal interface. Calculated as the difference in peak-to-peak values in the Carmel Formation lithologic units. Defined by changing GR log values. Bottom: Differences in dynamic Young’s modulus taken between average values of lithologic units in the Carmel Formation.
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CHAPTER 3

Identification of Reactivation and Increased Permeability Associated with a Fault Damage Zone Using a Multidisciplinary Approach

Abstract

We evaluate the fault damage zone associated with a reactivated long-strike length, small-offset normal fault in the Permian Cedar Mesa Sandstone, southeastern Utah. This fault is characterized by a single slip surface and a 9-m wide damage zone containing deformation bands and veins. Field observations include cross-cutting relationships, permeability increase, rock strength decrease, and ultraviolet-light-induced mineral fluorescence within the damage zone. These field observations, combined with the interpreted structural diagenetic sequence from petrographic analysis, suggest a deformation history of reactivation and several generations of mineralization. All deformation bands and calcite veins fluoresce under ultraviolet light, indicating fluid pathway connectivity and a shared mineralization history. Preexisting structures act as loci for younger deformation and mineralization events, so this fault and its damage zone illustrate the importance of the fault damage zone to subsurface fluid flow.

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1Paper published with co-authors as listed:
ES Petrie*, RA Petrie§, JP Evans*
*Department of Geology, Utah State University, Logan, UT 84322-4505
§Baroid Industrial Drilling Products, Logan, UT 84341
We model a simplified stress history to understand the importance rock properties and variations in differential and effective mean stress have on the structures within the damage zone. The moderate confining pressures, possible variations in pore pressure, and porous, fine-grained nature of the Cedar Mesa Sandstone produces a fault damage zone characterized by enhanced permeability and mineralization.

Introduction

Fault-zone architectural components and their associated effects on permeability act as a primary control on fluid flow in the upper brittle crust (Knipe, 1993; Mozley and Goodwin, 1995; Caine et al., 1996; Evans et al., 1997; Rawling et al., 2001). Fault-zone architecture refers to the nature and geometry of the fault core, damage zone and protolith, which can be lithologically heterogeneous and structurally anisotropic (Caine et al., 1996). Spatial variations in the permeability of the fault damage zone, and fault core create conduits and/or barriers for fluid flow in faulted rocks (Aydin, 1978; Aydin and Johnson, 1978; Antonellini and Aydin, 1994; Aydin and Eyal, 2002; Lothe et al., 2002; Davatzes et al., 2003; Johansen and Fossen, 2008). Understanding sub-seismic fault damage zones is critical to understanding fluid flow in oil and gas production and sequestration operations for nuclear and chemical waste storage or CO2 sequestration scenarios. Consequently, we examine a small-offset, long-strike length normal fault so as to
characterize a field example of a damage zone with evidence for multiple
generations of mineralization due to dilatancy at moderate confining pressures in a
fine-grained porous sandstone.

In this paper, we examine the mineralization of a fractured fault damage zone
and propose a stress history using a uniaxial strain model in a normal Andersonian
tectonic stress orientation. The damage zone of this fault provides evidence for the
localization of stress and re-activation of pre-existing structural discontinuities,
as well as increased permeability and subsurface fluid flow. We use traditional field
and petrographic investigative methods, and introduce a novel method of ultra-
violet light illumination to document the presence and connectivity of veins within
the fault damage zone. We describe a diagenetic sequence and propose a stress
history based on rock properties, which supports failure in a dilational regime,
resulting in enhanced permeability and subsequent mineralization within the
damage zone.

Geologic Setting

Located on the White Canyon Slope in the upper Lake Powell region, SE Utah,
are several steeply dipping (78° to 90°), northwest-striking normal faults near Hite
Crossing, Utah (Figs. 3-1, 3-2 and 3-3). Accessible outcrop exposures lie in Glen
Canyon National Recreation Area near the confluence of the Dirty Devil and
Colorado Rivers. The faults cut the Permian Organ Rock Formation and the
underlying Cedar Mesa Sandstone of the Cutler Group (Willis, 2012). The Organ Rock Formation is 60-80 m (196-262 ft) thick in the study area and consists of reddish-brown interbedded siltstone, mudstone and fine to medium-grained sandstone (Willis, 2012). The underlying Cedar Mesa Sandstone is a light-gray to orange, cross-bedded, fine-grained quartz arenite with an overall thickness of 300-365 m (1000-1200 ft) bed sets within the Cedar Mesa Sandstone range from 1-15 m (3.28-49 ft) thick. Deposition of the Cedar Mesa Sandstone occurred dominantly within eolian systems but interbedded siltstones represent depositional environments associated with transient streams, small rivers, and/or floodplains (Loope, 1985; Langford and Chan, 1988; Anderson et al., 2003; Willis, 2012).

At this field locality, the Permian sediments were deposited on the margins of the Paradox Basin and are not affected by subsurface salt movement. The field locality is surrounded by Laramide-related monoclines such as Comb Ridge, the Waterpocket Fold, and the San Rafael Swell, and Cenozoic laccoliths of the Abajo Mountains to the west and the Henry Mountains to the east (Fig. 3-1). The normal fault evaluated in this study has an estimated strike length of up to 25 km, determined from geologic maps and air photos. The west-northwest (287°-295°) strike is consistent with the modern stress orientation in the area and with the Four Corners structural lineament (Baars and Stevenson, 1982; Heidbach et al., 2008). Estimated vertical offset, as observed from stratigraphic offset between the Cedar Mesa Sandstone and overlying Organ Rock Formation and White Rim Sandstone is approximately 30 m within the study area (Fig. 3-1).
Figure 3-1. A) Location and tectonic elements of the Colorado Plateau, modified from Kelley, 1955 and Condon, 1997 with geologic base map of the study area modified from Willis, 2012. B) Google Earth imagery of study area showing landscape-scale bleaching in the Organ Rock Formation and location of fault damage zone with UV-light induced fluorescence evaluated in this study, labeled (UV). Note bleaching along fault traces within the Organ Rock Formation.
Although the vertical offset is small, diagenetic alteration patterns and mineralization are observed in outcrop including veins and cements that highlight the nature of the fault damage zone (Figs. 3-1 and 3-2; Willis, 2012). Also, landscape-scale alteration and fracture adjacent bleaching, via alteration of iron creating yellow to tan fracture margins is associated with the faults cutting the Organ Rock Formation in both the field and on remote imagery (Figs. 3-1 and 3-2; Appendix Table C-1). The Organ Rock Formation exhibits mineralized fractures with bleached margins that parallel the structural orientation of discontinuities in the underlying Cedar Mesa Sandstone. The outcrop study locality is near the southeastern tip of a normal fault, and in map pattern the fault is curvilinear and associated with several synthetic and antithetic branch faults (Figs. 3-1 and 3-2). All faults, joints, and veins dip steeply (78°-90°; NE / SW) (Fig. 3-3; Willis, 2012).

Methods

Data collected at the outcrop location include measurement of fracture orientation and spacing, examining the presence and nature of fracture-associated mineralization, as well as field-derived compressive strength and permeability data. These field data are used in combination with digital photographs of mineral fluorescence taken at night under ultraviolet light (UV) to characterize the fault damage zone. Details of the camera specifications and UV light source are in Appendix B-2. Mineral fluorescence is a phenomenon that occurs when minerals
Figure 3-2. Geologic map of study area, 1:7,500, photos of the exposed fault zone in the Cedar Mesa Sandstone and overlying Organ Rock Formation. Note the hematite staining associated with the Cedar Mesa Sandstone fault and alteration associated with faults crossing the Organ Rock Formation. Slickenlines on fault plane indicate down to the south dip-slip.

Irradiated with ultraviolet radiation re-emit light in the visible range. The irradiation raises the energy level of electrons displacing them to a higher energy level within an atom (Wenk and Bulakh, 2004). As the electrons return to the ground state, they release energy in the form of visible light (Wenk and Bulakh, 2004). Scanline fracture inventories (LaPointe and Hudson, 1985; Priest, 1993) are combined with permeability and compressive rock-strength transects to identify
changes in these rock properties across the fault zone. An N-type Schmidt hammer was used to collect estimates of outcrop-based compressive strength along the scanline. The Schmidt hammer was oriented vertically to the bedding plane and rebound values were recorded on average every 3 m along the scanline transect. Detailed methodology for these measurements is given in Selby (1993), and Chapter 2. Outcrop air-permeability data were collected vertically at the same locations as the Schmidt hammer data using a TinyPerm II field permeameter (NER, 2011). The outcrop-derived air permeability and compressive strength measurements were obtained on clean surfaces away from cracks or edges to limit the effects of adjacent free surfaces. These data likely overestimate permeability by a factor of ~1.7 (Fossen et al., 2011), underestimate compressive strength (Selby, 1993) and should not be extrapolated to assumed subsurface conditions, but do provide outcrop-based data essential in delimiting variations in mechanical properties associated with the spatial distribution of the fault zones and host rocks. Details of the Schmidt Hammer and TinyPerm II specifications and their use at the outcrop are in Appendix B-2.

Petrographic and geochemical analyses were conducted using thin sections, powder X-ray diffraction, and 55-element analysis through sodium-peroxide-fusion inductively coupled plasma mass spectrometry and atomic-emission spectroscopy (ICPM90A) of the whole-rock and vein mineral samples (Appendix C-2). Hand samples include off-fault sandstone protolith as well as samples from the damage zone. The damage zone samples represent the host-rock sandstone and associated
veins. In this paper, we refer to the undamaged off-fault Cedar Mesa Sandstone as the protolith, and the damage zone and fault adjacent sandstones that contain open joints, deformation bands, and / or veins as the host-rock. Joints in this study are open discontinuities generally occurring as large lineaments. Veins are mineralized fractures, cemented with calcite and/or hematite.

Host-rock samples with veins were trimmed using a rock saw or Dremel drill to separate vein minerals from the host-rock sandstone prior to petrographic and geochemical analysis. Optical thin-section analysis was done using a Zeiss polarizing petrographic microscope. Photomicrographs were imported into ImageJ for quantification of grain size and porosity analysis. Powdered samples for ICPM90A analysis were sent to the commercial laboratory SGS, Vancouver, BC.

The diagenetic history defined in this paper includes the burial and structural diagenesis associated with the Cedar Mesa Sandstone, as interpreted from field and petrographic relationships. The interpreted structural diagenesis is used in conjunction with stress-history analysis to understand the deformation and mineralization events of this fault damage zone. Based on the structural diagenetic sequence we model a simplified stress history using Coulomb failure criteria and vary the coefficient of friction, m, for each deformation event. To best describe the stress path through time, we plot the modeled results from Mohr-Coulomb analysis on a q-p diagram to evaluate a likely history of mechanical failure and deformation within the Cedar Mesa Sandstone.
The q-p diagram is based on the same principle as that of a Mohr-Coulomb diagram but allows more data to be incorporated, including the effect of pore fluid pressure and changes in grain size (Wong et al., 1992; Schultz and Siddharthan, 2005). The axes of the q-p diagram are defined by q, differential stress, and p, effective mean stress, the yield surface which separates the inelastic from elastic regimes is an elliptical envelope that intersects the p axis at point P*, the grain-crushing pressure, and the critical state line, which describes the coefficient of internal friction (Wong et al., 1992; Schultz and Siddharthan, 2005). The intercept at point P*, (Fig. 3-4) scales with changes in porosity, (\(\phi\)) and average grain size, (R), where

\[ P^* = (\phi R)^{-1.5}. \]  

We use porosity and grain size from thin-section analysis to calculate P* and create yield envelopes for the q-p diagram. P* values shown in figure 3-4 are based on thin-section porosity and grain size values within the protolith and host-rock sandstones away from veins and vein cementation.

Results

Field Observations

At the landscape-scale the faults in the Hite fault array are easily identified within the Organ Rock Formation siltstone and mudstone by changes from red-brown to yellow-tan due to a decrease in iron concentrations of up to 30 weight
percent (Figs. 3-1 and 3-2; Appendix Table C-1). These color changes surround the fault traces, are often symmetric along fracture margins, vary from 5 cm to 2 m in width, and are associated with increased fracture densities (Fig. 3-2). At the map scale (1:7,500), the fault trace is curvilinear and as it nears the fault tip has a horsetail splay of antithetic and synthetic faults (Fig. 3-2). Within the Cedar Mesa Sandstone structures in the fault damage zone include deformation bands, hematite- and calcite-filled veins, shear fractures, and open joints. At the outcrop the main slip surface of the fault strikes northwest, and the mean orientation is 304°/81°S (±4°).

Slickenlines on the fault plane indicate dominant dip-slip down to the south (Fig. 3-2).

Figure 3-4. Schematic q-p diagram, definition of yield envelopes based on mean grain size observed in thin-section. P* in undamaged protolith is 252 MPa; P* for damage zone host-rock is 122 MPa and 155 MPa. The intercept at point P* scales with changes in porosity, (f) and average grain size, (R), where $P^* = (\phi R)^{-1.5}$. 
Deformation bands occur in two orientation sets: the older set, (DB1), has a strike orientation to the north (020°) and the younger set, (DB2), is orthogonal to DB1 with a strike orientation to the west-northwest (303°) (Fig. 3-3). Deformation bands are limited in their distribution to the fault footwall, and occur in a cluster adjacent to the fault with a median distance between deformation bands of 0.085 m (Fig. 3-5). Fracture densities ranging from 6 to 13 fractures per meter and increased field-derived air permeability and decreased rock strength are observed within the fault damage zone (Figs. 3-5 and 3-6). Calcite veins are ubiquitous throughout the scanline survey, but occur in higher density clusters of 5 fractures per meter in both the hanging wall and footwall of the fault. Some calcite veins occur adjacent to or surround the hematite veins at their interface with the sandstone host-rock and are also found in the core of deformation bands (Figs. 3-5 and 3-6). Hematite veins occur dominantly in the hanging wall of the fault damage zone with a fracture density of 6 fractures per meter (Fig. 3-5). Hematite veins and zones of iron-oxide cementation occur adjacent to the fault and hematite occurrence decreases away from the fault (Fig. 3-5). Ellis et al. (2012) found a similar distribution of fracture patterns and fracture system evolution, to that observed in this study, including old deformation bands cross-cut by younger fracture sets, which appear in clusters but are not abundant in the entire rock volume. They proposed that clustering of fractures is a product of localized deformation, such as faulting or bed-parallel slip (Ellis et al., 2012).
Structural orientation data show that veins, faults, and open joints are steeply dipping (75-90°) and share a dominant northwest strike orientation similar to that of DB2 (Fig 3-3). Cross-cutting relationships show deformation band set 1 is offset by deformation band set 2, veins, and joints (Fig. 3-6). We consider the barren joints, which occur throughout the field area, and are often large-scale, to be the youngest structure.

Digital photographs taken of the pavement surface illuminated under a UV light source highlights the connectivity between fractures of all orientations and the extensive mineralization of the veins which is not obvious in visible light (Fig. 3-6). The yellow to white UV-induced fluorescence response occurs in the calcite vein fill and calcite-cored deformation bands of the fault damage zone, but such UV-induced fluorescence is absent in the protolith sandstones. The observed UV-induced fluorescence response is coincident with the occurrence of increased fracture densities (Fig. 3-6), increased air permeability, and decreased outcrop compressive strength (Fig. 3-7). Based on these field observations, we estimate that the fault damage zone is 9 m in width.
Figure 3-5. Fracture occurrence and type versus distance along Cedar Mesa Sandstone scanline. Purple bar labeled fluorescence shows distribution of UV-induced fluorescence response.
Figure 3-6. Daylight and UV light induced fluorescence photos, same position between photo pairs marked by red A/B. A) Distribution of UV-induced fluorescence associated with deformation band sets 1 (DB1) and 2 (DB2), area of red box enlarged in B. B) Offset of DB1 by DB2, note calcite vein in core of deformation bands. C) Mineralized calcite veins cross-cut zone of iron-oxide cement. H – hematite vein, C – calcite vein, Fe – zone of iron-oxide cement.
Figure 3-7. Compressive rock strength derived from Schmidt hammer and air permeability data along fracture-data scanline, zone of fluorescence shown with purple bar (Fig. 3-5).

Thin-Section Petrography

Cedar Mesa Sandstone Protolith Petrography

The Cedar Mesa Sandstone protolith is a medium-silt to medium-sand sized quartz arenite with an average grain size of 0.1 mm (range 0.04-0.36) (Table 3-1 and Fig. 3-8). The protolith contains rare calcite cement, rare hematite and calcite grain coatings, and long or tangential grain contacts (Fig. 3-8). The protolith sandstone has an estimated intergranular volume of 41%, which includes 26% open porosity, and 15% quartz overgrowth cement and calcite cement (Fig. 3-8 and Table
3-1) (Paxton et al., 2002). The protolith sandstone is light gray to tan-orange unlike at other localities in Utah, where it is red due to the presence of hematite grain coatings (Mountney and Jagger, 2004).

Cedar Mesa Sandstone Damage Zone Petrography

Optical petrography of the host-rock samples from the fault damage zone show similar QFL distributions as the protolith, dominantly quartz (90-95%), some feldspar (2-5%), and rare lithic clasts (<2%). Within the damage zone, intergranular volume ranges from 17-54%, with the greatest intergranular volume occurring adjacent to veins, where an abundance of calcite and/or hematite cement exists. In thin-section, vein aperture ranges from 0.07-2.01 mm and vein mineralogy includes calcite and hematite (Table 3-1).

Quartz overgrowth cement is common in host-rock samples, these quartz overgrowths often encase hematite grain coatings, (Fig. 3-9), and predate open-mode fractures as grains with quartz overgrowth are cut by hematite and calcite veins (Fig. 3-10). Based on the hematite psuedomorphs and the cross-cutting relationships observed between calcite and hematite veins, we propose several generations of vein mineralization (Figs. 3-9 to 3-12). Calcite veins show a minimum of two stages of precipitation with symmetric euhedral calcite crystals that generally line fracture walls and a core of mottled ferroan calcite (Fig. 3-11). Hematite often takes on a dogtooth-spar crystal form, where these psuedomorphs represent replacement of an early calcite or siderite phase by hematite (Fig. 3-10).
We note that the euhedral calcite veins tend to line fracture walls whereas the ferroan mottled calcite occurs as a fill between mineralized fracture walls at times preserving fracture porosity, and as such this cement may be postkinematic (Laubach, 2003). In thin-section all calcite veins fluoresce when exposed to transmitted and reflected UV light. Hematite veins are cross-cut by ferroan mottled calcite veins suggesting that this is the latest and perhaps final stage of calcite mineralization (C3) (Fig. 12).

Figure 3-8. Cedar Mesa Sandstone protolith photomicrograph. Q – quartz, F – feldspar, Fd–feldspar dissolution, P – pore space. See Table 3-1 for complete petrographic analysis.
Figure 3-9. Photomicrograph of host-rock of the Cedar Mesa Sandstone showing example cements, including calcite and quartz overgrowths. The quartz overgrowths preserve hematite grain coatings, dissolution of feldspar grains created intragranular porosity. Q – quartz, OG – quartz overgrowth, H – hematite, L – lithic, Fd – feldspar dissolution, C1 – calcite cement.
Figure 3-10. Photomicrograph of vein within the damage zone with hematite psuedomorphs after calcite or siderite, and microfractures filled with calcite and hematite. Q – quartz, OG – quartz overgrowth, H – hematite.
Figure 3-11. Photomicrographs from fault damage zone, with three types of calcite cement. Inset red box defines area of 10x magnification of photomicrographs shown below. C1 – calcite cement, C2 – euhedral calcite, C3 – ferroan calcite, P – pore space, Q - quartz.
Figure 3-12. Fault damage zone photomicrograph with mottled ferroan-calcite vein cutting hematite vein. Antitaxial hematite vein contains host-rock quartz grains, Q – quartz, H – hematite, C3 – mottled ferroan-calcite cement
Table 3-1. Petrographic data obtained from thin-section analysis; composition data is % whole rock. Grain size measurements and intragranular volume estimates made using ImageJ. These measurements may be overestimates of intragranular volume if grain plucking occurred during thin-section preparation. Grain size measurements are minimum and maximum lengths measured using photomicrograph analysis. Fracture-adjacent grain size measured to 0.5 mm on either side of fracture. UV activated fluorescence observed from both transmitted and reflected UV light source.

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Sample Type</th>
<th>Grain Size Range Median (mm)</th>
<th>Detrital Grains</th>
<th>Cement</th>
<th>Intragranular porosity</th>
<th>Vein Minerals Aperture (mm)</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>GCCM1</td>
<td>Protolith</td>
<td>0.04-0.36 0.1</td>
<td>61.6 0.8 2.0 9.6 11.2 0.0</td>
<td>14.8</td>
<td>N/A</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GCCM2</td>
<td>Fault damage zone</td>
<td>0.04-0.21 0.1</td>
<td>62.4 0.0 0.0 0.0 0.0 21.6</td>
<td>16.0 Hematite 0.32 Calcite* 0.05-0.07</td>
<td>Calcite vein cuts hematite vein at 81°</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GCCM3</td>
<td>Fault damage zone</td>
<td>0.04-0.23 0.1</td>
<td>66.4 0.0 0.0 5.6 16.0 3.6</td>
<td>8.0 Calcite* 0.7-0.8</td>
<td>Euhedral calcite with disseminated quartz clasts</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GCCM4</td>
<td>Fault damage zone</td>
<td>0.03-0.23 0.09</td>
<td>33.2 0.0 0.0 6.0 56.4 0.0</td>
<td>4.4 Calcite*</td>
<td>Euhedral calcite at fracture margins mottled calcite in center</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GCCM6</td>
<td>Fault damage zone</td>
<td>0.05-0.33 0.14</td>
<td>52.4 2.0 0.0 17.6 7.6 3.2</td>
<td>17.2 Hematite Calcite*</td>
<td>Fracture parallel calcite filled quartz-grain microcracks</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GCCM8</td>
<td>Fault damage zone</td>
<td>0.04-0.19 0.09</td>
<td>41.0 0.4 0.0 5.2 1.2 47</td>
<td>5.6 Hematite 0.25-1.32 Calcite* 0.07-0.13</td>
<td>Fracture parallel calcite filled quartz-grain microcracks Euhedral iron oxide after calcite pseudomorphs (Fig.3-10)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GCCM9</td>
<td>Fault damage zone</td>
<td>0.03-0.2 0.11</td>
<td>36 0.0 0.4 8.0 9.0 32.0</td>
<td>14.0 Hematite 1.05 Calcite* 0.14-0.48</td>
<td>Antitaxial hematite and calcite veins</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Trace Element Analysis

We examine the trace-element geochemistry of whole-rock samples from the sandstone protolith, the damage zone host-rock, and veins to verify the source of the UV fluorescence and identify any elemental enrichment in the fault damage zone. The presence of uranium in calcite is known to cause a yellow/white fluorescence when exposed to UV light (Wenk and Bulakh, 2004). Trace element analysis shows that in addition to uranium the veins are enriched in barium, copper, zinc, arsenic, colbalt, molybdenum, lead, antimony, and tungsten relative to global sedimentary rock crustal concentrations, the protolith, and host-rock samples (Fig. 3-13 and Appendix Table C-2) (Gabelman, 1977; Rudnick and Gao, 2003).

We find increased concentrations of uranium in all vein samples associated with the fault zone at 1.41 to 12.4 ppm versus 0.69 ppm in the protolith (Fig. 3-13 and Appendix Table C-2). Uranium concentrations in the sandstones adjacent to the fault damage zone range from 0.34 to 0.96 ppm while the calcite-filled fractures contain 1.41-1.93 ppm of uranium and the hematite veins contain 5.58 to 12.4 ppm of uranium. Arsenic concentrations are 465-650 times higher than that of the protolith and arsenic shows the greatest change in relative concentration between the sandstone protolith and the vein material. Elemental enrichment occurs within both the host-rock sandstone of the fault-damage zone and its associated veins, however, readily fluid mobile elements such as As, Ba, U, and W show highest enrichment within the veins (Fig. 3-13 and Appendix Table C-2).
Figure 3-13. Geospider diagram, double normalized against published values of continental crust sedimentary rock concentration (Rudnick and Gao, 2003), and immobile REE Nd.

Interpretation

Fracture Evolution And Damage Zone Permeability

We interpret the diagenetic and fracture sequence based on field and petrographic observations to understand the susceptibility of this fault damage zone to reactivation and fluid flow (Fig. 3-14). We document two stages of deformation band formation, the first set, DB1, is north-striking and offset in a right-lateral sense by the west to north-west striking deformation band set, DB2 (Fig. 3-6). The locally observed change in structural orientations between DB1 and the younger structures
such as DB2 suggests a local rotation of the horizontal stress orientation within
the damage zone (Faulkner et al., 2006). The dominant far-field west to north-west
structural orientations are consistent with modern stress orientations and with the
inferred regional stress orientation of the Four Corners structural lineament, which
suggests a long-lived and regionally dominant west to north-west orientation of
maximum horizontal stress (Baars and Stevenson, 1982; Heidbach et al., 2008).

The third phase of deformation is characterized by shear, extensional-shear,
and open-mode fractures that record a mineralization history of calcite and
hematite. The cross-cutting relationships and changing vein mineralogy associated
with phase III suggest a history of reactivation. We suggest that the presence of pre-
existing structural discontinuities provide a loci for the nucleation of deformation
through time, and in this scenario a zone of enhanced permeability. This
interpretation is supported by the observed cross-cutting relationships between
fracture sets, and their locally high abundance within the fault damage zone. The
UV-induced fluorescence is associated with fracture sets of all orientations but
limited in its occurrence to the fault damage zone where it highlights the
connectivity between fractures within the fault damage zone. We infer that this
distribution is consistent with a near simultaneous mineralization or a similar or
shared fluid source through a sequence of fracturing events (Figs. 3-5 and 3-7).
Figure 3-14. Interpreted diagenetic sequence for the Cedar Mesa Sandstone, showing the chronologic order and relative duration of diagenetic events. Order is based on field and petrographic observations of cross-cutting relationships and crystal form. Formation or reactivation of fracture sets ended with calcite or hematite mineralization. Grey bars are used to highlight uncertainty where timing is harder to establish.

Stress Path Analysis

Using the interpreted sequence of structural diagenesis, the deformation history for this fault damage zone is simplified to 3 phases. To understand the effect of the deformation history on permeability distribution in the fault damage zone, we model failure in the subsurface using a uniaxial-strain reference state and variations of the coefficient of friction, m, plotting the stress history on a q-p diagram (Fig. 3-15). We assume an Andersonian stress state for a normal fault ($S_v=\sigma_1$);

$$S_v=\rho gz$$  \hspace{1cm} (2)\]

$$S_H=S_h=\left(\frac{v}{1-v}\right) \rho gh$$  \hspace{1cm} (3)\]

<table>
<thead>
<tr>
<th>Diagenetic Sequence</th>
<th>Relative Timing</th>
</tr>
</thead>
<tbody>
<tr>
<td>Early</td>
<td>Late</td>
</tr>
<tr>
<td>Burial</td>
<td></td>
</tr>
<tr>
<td>Hematite grain coating</td>
<td></td>
</tr>
<tr>
<td>Compaction</td>
<td></td>
</tr>
<tr>
<td>Quartz overgrowths (OG)</td>
<td></td>
</tr>
<tr>
<td>Regional Bleaching</td>
<td></td>
</tr>
<tr>
<td>Calcite cements (C1)</td>
<td></td>
</tr>
<tr>
<td>Feldspar weathering (Fw)</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td>phase I</td>
<td></td>
</tr>
<tr>
<td>Deformation band set 1 (DB1)</td>
<td></td>
</tr>
<tr>
<td>phase II</td>
<td></td>
</tr>
<tr>
<td>Deformation band set 2 (DB2)</td>
<td></td>
</tr>
<tr>
<td>phase III</td>
<td></td>
</tr>
<tr>
<td>Open mode fractures</td>
<td></td>
</tr>
<tr>
<td>Shear fractures</td>
<td></td>
</tr>
<tr>
<td>Calcite veins (C2)</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td>Structural</td>
<td></td>
</tr>
<tr>
<td>Open mode fractures</td>
<td></td>
</tr>
<tr>
<td>Shear fractures</td>
<td></td>
</tr>
<tr>
<td>Hematite veins &amp; pseudomorphs</td>
<td></td>
</tr>
<tr>
<td>Iron-oxide cement zones</td>
<td></td>
</tr>
<tr>
<td>Open mode fractures</td>
<td></td>
</tr>
<tr>
<td>Calcite veins (C3)</td>
<td></td>
</tr>
<tr>
<td>Regional joint-set</td>
<td></td>
</tr>
</tbody>
</table>
(Anderson, 1951; Engelder, 1993). This initial state of stress is derived using an estimated maximum burial depth of 2 km, consistent with stratigraphic thickness in the area (Hintze and Kowallis, 2009), and is uncorrected for compaction. We use small variations in the coefficient of friction, where

\[ \mu = \tan \Phi, \]

\[ (4) \]

from an initial value of \( m = 0.6 \). Three phases of deformation are modeled, two of which created deformation bands, followed by a third phase of open-mode fracture development and vein mineralization.

In all the models, we assume a cohesive strength of 15 MPa and vary the angle of internal friction (\( \Phi \)) from 30° to 36°. The variation in \( F \) depends on the acquisition of structural discontinuities through time because the presence of structural discontinuities can serve to increase or decrease the internal friction locally depending on the discontinuity type (Anderson, 1951; Byerlee, 1978; Underhill and Woodcock, 1987; Mair et al., 2000; Li et al., 2005; Kaproth et al., 2010) (Fig. 3-15).

Our models assume that the development of deformation bands increases the coefficient of friction by strengthening the material relative to the parent material via shear-driven cataclasis or denser grain packing (Underhill and Woodcock, 1987; Li et al., 2005; Kaproth et al., 2010). Each deformation phase represents changes in differential stress and/or pore fluid pressure conditions, and are plotted on a normal stress/shear stress diagram. When the Mohr circle crosses the Coulomb failure envelope, which is defined by a specific \( \Phi \) value, failure occurs.
Phase I deformation of the parent material occurred during increased differential stress from the initial stress state, producing the deformation band set 1 (DB1). For this phase, when the differential and effective mean stress values are plotted on the q-p diagram, failure is shown to have occurred in the dilational regime (Fig. 3-15).

Phase II deformation generated deformation band set 2, which offset DB1 (Fig. 3-6 A and B). We evaluate this phase using $\mu=0.73$ ($\Phi=36^\circ$), based on the assumption that the friction angle increased due to presence of DB1 and associated local material strengthening (Fig. 3-15) (Underhill and Woodcock, 1987; Mair et al., 2000; Li et al., 2005). To initiate failure under these conditions, differential stress would have increased, triggering mechanical failure that is represented by a point in the dilational regime of the q-p diagram (Fig. 3-15). During phase I and II, the necessary differential stress for failure and formation of deformation bands is consistent with experimentally determined differential stresses for the formation of deformation bands, which reach a yield point of up to 125 MPa (Mair et al., 2000).

The final phase of deformation, phase III, we believe that the value of $\mu=0.73$ ($\Phi=36^\circ$) is still applicable as the host rock now contains two sets of deformation bands. Phase III deformation generated open-mode fractures and the mineralization of the pre-existing deformation bands. Because of the formation of veins and mineralization of pre-existing structures during phase III we evaluate failure during phase through increased pore-fluid pressure and at lower differential stress conditions (Fig. 3-15). Formation of open-mode fractures have been shown
to occur and propagate under small strain load conditions (Olson et al., 2009) (Fig. 3-15). In phase III increased pore-fluid pressure induces mechanical failure to occur for a reduced mean effective stress and resulting in mechanical failure represented by a point in the dilational regime of the q-p diagram.

Figure 3-15. Q-P diagram modeling simplified deformational history of the Cedar Mesa Sandstone and the calculated stress path for the Cedar Mesa sandstone, with Mohr-Coulomb inset for reference. Color of lines representing angle of internal friction in Mohr-Coulomb inset and the critical state line on the q-p diagram represent the same $\Phi$ value. Color and pattern of Mohr circle in Mohr-Coulomb inset corresponds to stress path line on q-p diagram for deformation phases I-III.
Implications for Stress State and Deformation Sequence

In the brittle realm, rock strength increases with burial depth and is dependent on fluid pressure. Absolute strength however is dependent on lithology and the presence and distribution of existing structural discontinuities (Byerlee, 1978; Sibson, 1985; Gale et al., 2007; Fossen, 2010). The interplay of these intrinsic rock strength parameters and extrinsic forcing factors can lead to reactivation of existing faults and fractures, including mineralized veins and open joints. This reactivation can occur at much lower levels of stress than needed for the intact host rock (Byerlee, 1978; Gale et al., 2007; Fossen, 2010).

Previous researchers have evaluated the reactivation potential of existing planes of weakness in several ways (Sibson, 1985; Morris et al., 1996; Tong and Yin, 2011; Leclère and Fabbri, 2013). These evaluations use Coulomb failure criteria; assume a uniform state of stress, and negligible or cohesionless materials with no variation in the coefficient of friction (Sibson, 1985; Morris et al., 1996). These scenarios show that reactivation can occur at low differential stress conditions, as $s_3$ tends toward zero, due to increased pore fluid pressure / overpressure.

The Mohr-Coulomb failure analysis of the three deformation phases modeled for this fault damage zone examines the differential and effective mean stress values required for failure. When plotted on a q-p diagram, the differential and effective mean stress points represent a hypothetical stress magnitude path within the fault damage zone (Fig. 3-15). In this simplified model of stress magnitude history and rock properties, cohesive strength is held constant but the coefficient of friction
varies and failure occurs consistently within the dilational region. In both phase I and II, deformation occurs under higher differential stress and effective mean stress conditions than that of the previous stress state, i.e. the initial state of stress and phase I respectively. During phase III, deformation occurs for increased pore fluid pressure conditions that weakens the rock, allowing mechanical failure under lower differential and effective mean stress conditions, and reactivation of structures in a locally stronger material relative to the parent material.

The stress history model presented here is consistent with the field and laboratory observations of the structures found in this fault damage zone. These structures resulted in enhanced subsurface permeability that allowed fluid flow and mineralization to occur within the damage zone. Due to the fine-grained and porous nature of the Cedar Mesa Sandstone, combined with the estimated initial stress state, the modeled stress history never exceeds the necessary pressure gradients to create pure compaction bands (Fig. 3-15). These dilational features, expressed as mineralized deformation bands or veins, act as transient permeable fluid pathways in the subsurface.

Conclusions

We interpret field and laboratory data that characterize fault damage zone features associated with a steeply dipping, small-offset, normal fault in the aeolian quartz arenite Cedar Mesa Formation. The 9 m wide fault damage zone consists of a
fault core dominated by a single slip surface and micro-scale cataclasis, and lacks any gouge or mylonite. The damage zone includes deformation bands, faults, veins, and joints clustered near the main fault slip-surface. Cross-cutting relationships and mineral replacement features within the damage zone are used to reconstruct the history of structural reactivation and mineralization.

We model the progressive formation of these structures based on the interpreted structural diagenetic sequence and stress history. Plotting the stress history in q-p space shows that theses structures formed in the dilational regime. The moderate confining pressures and porous, fine-grained nature of the Cedar Mesa Sandstone produces a fault damage zone characterized by enhanced permeability and mineralization. The modeled stress history presented here highlights the importance rock properties and variations in differential and effective means stress have on the structures within the damage zone.

Ultra-violet light induced fluorescence documents mineralization associated with extensional-shear reactivation of deformation bands, open-mode fractures and fracture connectivity within the fault damage zone. All deformation bands, structural features typically thought to be low permeability, and calcite veins in the damage zone fluoresce under ultraviolet light, indicating fluid pathway connectivity and a shared mineralization history. The UV-light response shows that the pre-existing structures act as loci for younger deformation and mineralization events. This fault example illustrates the importance of fault damage zone architecture and its effect on fluid flow and mineral distribution in the subsurface.


CHAPTER 4

Failure of Caprock Seals as Modeled from Mechanical Stratigraphy, Stress History and Tensile Failure Analysis of Exhumed Analogs.¹

Abstract

The varied sedimentologic and tectonic histories of clastic caprocks and their inherent mechanical properties control the distribution and morphology of permeable fractures. The migration of liquid or gas through mm- to cm-scale fracture networks can result in focused fluid flow and may compromise seal integrity. In order to understand the nature and distribution of fluid-flow pathways in caprock seals we examine four Paleozoic and Mesozoic analog outcrops in Utah that show evidence for subsurface fluid flow through permeable fracture networks. We combine outcrop analysis with subsidence analysis, paleo-loading histories, and rock strength testing data in uniaxial strain models to evaluate the likely timing of fracture initiation and fracture type.

Relative to the underlying sandstone reservoirs, all four seal types are low-permeability relative to the underlying reservoir rock (9.87x10⁻¹⁶ m² to 1.18x10⁻¹³ m² / 0.001 to 0.12 Darcy). These heterolithic sequences also show mineralization of

¹ Manuscript accepted AAPG Bulletin with the following co-authors:
ES Petrie* JP Evans* and SJ Bauer§
*Department of Geology, Utah State University, Logan, Utah 84322
§Geomechanics Research Center for Experimental Geosciences, Sandia National Laboratories, Albuquerque, NM, 87123
natural hydraulic extension fractures and slip planes of shear fractures. Burial-history models suggest that the caprock seal analogs reached a maximum burial depth of greater than 4 km and experienced a lithostatic load of up to 110 MPa (15,954 psi). Median tensile strength from indirect mechanical tests range from 2.3 MPa (333.6 psi) in siltstone to 11.5 MPa (1,667.9 psi) in calcareous shale. Analysis of the pore-fluid factor ($\lambda_v$) through time shows changes in the expected failure mode, from extensional shear to hydraulic extension, and the dependence of failure mode on a combination of mechanical rock properties and differential stress. With increasing lithostatic load, the amount of excess pore pressure above the hydrostatic gradient that is required to induce failure increases but is also lithology dependent.

Introduction

Fractures may control the location and volume of fluid flow in the subsurface and are important to a range of geo-engineering scenarios compromising naturally occurring fluid/gas traps, or engineered waste storage systems (Nelson, 2001). The presence of discontinuities in caprocks affects their mechanical and hydro-geologic properties; migration of fluids or gas through mm- to cm-scale discontinuity networks can lead to the failure of hydrocarbon traps or waste repositories (Ingram et al., 1999; Almon and Dawson 2004; Pruess, 2008). This paper evaluates the nature and mechanical behavior of caprock seals. We examine the mechanical and
fracture stratigraphy of Paleozoic and Mesozoic failed seal analogues and provide a comparison of four seal types exposed in central and south-eastern Utah on the Colorado Plateau (Figure 4-1). Each locality has experienced Sevier and Laramide shortening, normal faulting above dissolving salt, and more recently, late Cenozoic extension within or along the margins of the Colorado Plateau.

Figure 4-1. Location map of outcrops and schematic tectonic elements. Map modified from Kelley, 1955; Kelley and Clinton, 1960; Hintze et al., 2000. Study localities: 1 – Organ Rock Formation, 2 – Carmel Formation, 3 – Earthy Member Entrada Sandstone, 4 – Tununk Member Mancos Shale.
Owing to their poor outcrop preservation, sealing lithologies are often difficult to study in the field. We take advantage of these arid region exposures and employ varied field methods to measure in-situ properties, and use laboratory based measurements of rock yield properties to quantify the nature of the seals. We use outcrop surveys of discontinuity distribution and stratigraphic changes to identify the relationships between composition, structural diagenesis/paragenesis and loading history. Field datasets include occurrence of structurally associated diagenetic alteration, fracture distribution, morphology of fractures, as well as rock strength and permeability surveys. These outcrop datasets allow the evaluation of the nature, occurrence, and distribution of fluid flow pathways in the various sealing lithologies. Laboratory rock testing data provide tensile strength for estimation of the failure envelopes associated with each rock type. We combine these data with derived estimates of changes in differential stress through time from analysis of the burial and tectonic history of each locality to understand the link of seal failure to stress history.

Fracture formation, propagation, and distribution patterns are controlled by lithologic anisotropy and structural history (Woodward and Rutherford, 1989; Nelson, 2001; Laubach et al., 2009). The stress history, burial-deformation-uplift path, pre-existing fabrics/anisotropy, and structural position are critical to understanding fracture propagation and distribution. Lithologic controls include the mechanical properties associated with rock type, stratigraphic layering, porosity, and fluid content (Rijken and Cooke, 2001; Larsen et al., 2010).
Geologic Setting

The Paleozoic and Mesozoic rocks evaluated in this study experienced variable stress histories generated by their structural and exhumation histories and have varied rock properties due to lithologic changes inherited during deposition. The exposed caprock seal analogs evaluated in this study include the Permian Organ Rock Formation, Jurassic Carmel Formation, Jurassic Earthy Member of the Entrada Sandstone, and the Cretaceous Tununk Member of the Mancos Shale (Figure 4-2). The depositional setting for each seal analog is summarized briefly below.

Organ Rock Formation

Pennsylvanian and Permian sedimentary rocks, including the Organ Rock Formation, of the Cutler Group, were deposited within the restricted Paradox Basin. These units include marine limestone, shale, and evaporites, which filled the Paradox Basin with sediment derived from the Uncompahgre uplift of the Ancestral Rockies (Nuccio and Condon, 1996; Bump and Davis, 2003; Davatzes et al., 2003).

At its exposure near the confluence of the Dirty Devil and Colorado Rivers in Glen Canyon National Recreation Area, the Organ Rock Formation is cut by a NW-striking fault array that exhibits evidence for sub-surface fluid flow (see Chapter 3). The Organ Rock Formation is approximately 120 m (393.7 ft) thick, and is a sequence of interbedded mudstone, siltstone and very fine sandstone, interpreted to represent fluvial to marginal marine depositional environments. The Organ Rock
Formation conformably overlies the Cedar Mesa Sandstone reservoir analog, and coarsens upward into the overlying White Rim Sandstone (Figure 4-2).

Carmel Formation

Sedimentation during the Triassic and Jurassic was influenced by the development of magmatic arcs to the west that provided sediment through episodic uplift of source terrain (Dickinson and Gehrels, 2003; Dickinson, 2004). The Jurassic San Rafael Group, which includes the Carmel Formation, was deposited in a trough-like depression overlying the J-2 unconformity (Pipiringos and O'Sullivan, 1978). The Carmel Formation thickens and marine limestone facies dominate to the west, whereas to the east it thins and is siliciclastic rich (Hintze and Kowallis, 2009). The outcrop locality of the Carmel Formation occurs in the San Rafael Swell, an east-vergent, NNE-trending, doubly plunging anticline (Figure 4-1) (Bump and Davis, 2003). The San Rafael Swell is estimated to have formed between 93 Ma and 58 Ma and is associated with the Laramide Orogeny (Fouch et al., 1983; Lawton, 1985; Molenaar and Cobban, 1991; Guiseppi and Heller, 1998; Shipton and Cowie, 2001). Within the San Rafael Swell the Carmel Formation is a mixed siliciclastic-carbonate unit deposited in shallow marine to sabkha environments (Caputo, 2003; Hintze and Kowallis, 2009). We examine an exposure of the Carmel Formation along I-70, where the basal 39 m (128 ft) of quartz-rich pellooidal, micrite to packstone interbedded with fissile argillaceous-rich shale and gypsiferous sandstone are well exposed (Figure 4-2).
Figure 4-2. Measured stratigraphic columns from each outcrop locality shown in figure 4-1. Mean fracture density (number of fractures per meter along scanline transect) through the stratigraphic section is shown in bar graph. Black bars show scanlines adjacent to normal faults with > 1m (3.28 ft) offset.

Earthy Member of the Entrada Sandstone

The Jurassic Entrada Sandstone contains several members (Hintze and Kowallis, 2009), this study evaluates the low permeability Earthy Member exposed at Salt Wash Graben (Figure 4-1). The Earthy Member was deposited in a shallow
water marine environment and may represent reworking of the underlying Slick Rock Member (Otto and Picard, 1975). At the study locality, the upper sealing portion of the Earthy Member is 30 m (98 ft) thick and is characterized by interbedded siltstone and mudstone layers that overly a thickly bedded sandstone sequence below (Figure 4-2). This outcrop locality provides an opportunity to evaluate an exposed reservoir/seal interface as the partial exposure of an eolian interbed that has been bleached by a fluid in the subsurface prior to exhumation. This bleaching is thought to be due to reduction and mobilization of iron and then possible replacement with other minerals in the paleo-reservoir (Pearce et al., 2011).

Tununk Member of the Mancos Shale

The Tununk Member of the Cretaceous Mancos Shale is well exposed near Crystal Geyser, adjacent to the Green River (Figure 4-1). The Tununk Member was deposited in a deep-water, open marine setting of the Western Interior Seaway and is calcareous and sulfur-rich with ample forminifera and ostracod assemblages (Hettinger and Kirschbaum, 2002).

The outcrop is exposed in the hanging wall of the northernmost strand of the WNW-trending Little Grand Wash fault zone (Figure 4-1). The Little Grand Wash fault zone is a set of parallel normal faults with up to 210 m (689 ft) of throw (Shipton et al., 2005). The Mancos Shale is considered to be a regional seal and the Tununk Member consists of fissile lime-mudstone overlain by nodular
foraminiferous lime-mudstone. Bed thickness of the nodular lime-mudstone varies from 0.1 to 0.5 m (0.3 to 1.6 ft) (Figure 4-2, Table D-1); total regional thickness estimates of the Tununk Member ranges from 91 to 198 m (300 to 650 ft) (Schamel, 2006).

Methods

The four caprock seals that evaluated in this paper show evidence for the development of fracture permeability, fluid flow, and mineralization in the subsurface. For control, we compare the measured fracture distribution data to that of structural discontinuity data from core, where available, or to offset outcrop locations. Fracture spacing from core is estimated using the procedure set out in (Narr, 1996). These control localities are used to identify the effect of exhumation on fracture formation due to cooling and removal of overburden versus that of in-situ subsurface deformation. Data collected at each outcrop locality includes fracture orientation, spacing, termination, length, as well as the nature and distribution of mineralization following (La Pointe and Hudson, 1985; Nelson, 2001). Additionally, rock compressive strength was measured in the field with an L-type Schmidt Hammer, and air permeability data were obtained using TinyPerm II, (NER, 2011), where appropriate. Due to their fine-grained nature some of these rocks are below the detection limits of the Schmidt Hammer or TinyPerm II. The air permeability and compressive strength measurements were carried out on cleaned
surfaces away from cracks or outcrop edges to limit the effects of adjacent free surfaces. These data likely overestimate permeability (Fossen et al., 2011; Raduha, 2013), and underestimate compressive strength (Selby, 1993), but provide datasets for comparative relative analysis of rock strength and air permeability within the outcrops. These field data are used in combination with the quantitative stratigraphic analysis of fracture density distributions (Bertotti et al., 2007) from scan line fracture inventories (La Pointe and Hudson, 1985) and digital outcrop orthoimages to establish the variation in fracture distribution with changes in rock properties. The aim of combining criteria similar to those used in rock-mass strength ratings with the stratigraphic analysis of fracture distribution is to identify areas within the seal where changes in lithology result in significant changes in deformation behavior.

Petrographic analysis is used to identify relationships between diagenetic alteration and structural history. Optical petrography of thirty-eight thin sections was carried out using a Zeiss polarizing petrographic microscope. Photomicrographs were imported into JMicroVision (Roduit, 2013) for grain-size measurements and point-counting of detrital grains for mineralogic composition and estimation of pore volume. Representative samples of host rock, altered fracture-adjacent rock, and fracture-corridor samples were analyzed using powder X-ray Diffraction (XRD), and 54 elemental analysis through sodium peroxide fusion inductively coupled plasma mass spectrometry and atomic emission spectroscopy (ICPM90A) of whole-rock and vein-mineral samples. The x-ray diffraction analysis
was carried out at Utah State University. The rock powders were analyzed using XRD on a Panalytical diffractometer, using CuKa radiation, with angular variations from 2° to 72° 2Θ at 1°/min. A subset of samples, representative of host rock, alteration zone, and vein-minerals were used in whole-rock elemental analysis by SGS laboratories British Columbia.

Tensile strength tests of intact protolith rock samples from each locality were carried out by Sandia National Laboratories Geomechanics Research Department as part of a larger effort to constrain the mechanical behavior of seals. Tensile strength was determined via an indirect tensile strength test ASTM D 3967-05. Core samples with a 2.54 cm diameter (1 inch) were drilled from outcrop hand samples. The core samples were sliced perpendicular to the long axis in segments ~ 1.27 cm, (0.5 inches) in length; a rock length to diameter ratio of 0.5 was maintained for each sample. Each sample, at least five of each lithology were tested, was diametrically loaded between rigid platens (with bearing strips), until failure. The splitting tensile strength is calculated as follows:

\[ \sigma_t = \frac{2P}{\pi LD} \]  

where: \( \sigma_t \) is the splitting tensile strength (psi), \( P \) is maximum applied load, \( L \) is thickness of the specimen, \( D \) is the diameter of the specimen.

Burial stress history diagrams were created using compiled stratigraphic sections for the region of the outcrop localities (Hintze and Kowallis, 2009), uncorrected for compaction (Figure 4-3A). The burial history curves presented here
include an estimated thickness for younger-eroded sedimentary units (Cretaceous and Tertiary) based on regional sedimentary thickness patterns/distributions (Hintze and Kowallis, 2009). We use the burial history curves to consider the change in magnitude of the principal stresses through time as depth \((z)\) increases. We assume a uniaxial strain model reference state and Andersonian tectonic stress orientation for a normal fault regime \((S_v=\sigma_1)\), where:

\[
S_v = \rho g z
\]  
\[
S_H = S_h = \sigma_2 = \sigma_3 = \left(\frac{\nu}{1-\nu}\right) (S_v - P_p) + P_p
\]

\(\rho\) is rock density \((\rho=2.6 \text{ g/cm}^3 / 0.0939 \text{ lb/in}^3)\), \(g\) is the gravitational acceleration \((g=9.823 \text{ m/s}^2 / 32.174 \text{ ft/s}^2)\), \(\nu\) is Possion’s ratio \((\nu=0.25)\), and \(P_p\) is hydrostatic pore fluid pressure (Anderson, 1951; Eaton, 1969; Engelder, 1993). The stress history models are based on the results from the combined burial history created in OSXBackstrip (Cardozo, 2010) and Mohr circles with Griffith-Coulomb failure envelopes graphic plots created in GeoGebra (Hohenwarter, 2013). Using inflection points identified along the burial history curves at depths greater than 0.5 km Mohr circles are created for the specific time points (Figure 4-3A). Rock properties used to plot the Mohr circles and Coulomb failure envelopes were assumed to have a coefficient of internal friction, \(\mu=0.6\) and for intact media the cohesive strength \((C)\) was estimated from the results of the indirect tensile strength test given that:

\[
T = \frac{C}{2}
\]

(Griffith, 1924; Jaeger et al., 2007; Gudmundsson, 2011). We also consider a cohesionless material case where \(C=0 \text{ MPa} (0 \text{ psi})\) (Figure 4-3B).
Figure 4.3. (A) Example of burial history model used to consider changing stress magnitude through time. (B) Depth points from burial history curve are used to estimate $\sigma_1=S_v$ and $\sigma_3=S_h$ for Mohr circle. The Mohr circle is used to estimate the increase in pore fluid pressures required for failure at cohesive strength values $C=0$ and $C=6.3$ MPa (913.74 psi) and coefficient of internal friction of 0.6.
Field Observations

Mudstone and Siltstone - Organ Rock Formation

Bedding within the Organ Rock Formation dips gently (12° ± 2° W) and is cut by steeply dipping mineralized fractures. These fractures include shear fractures with bleached fracture margins and mineralized veins distributed throughout the entire stratigraphic thickness, and occur adjacent to normal faults (Figure 4-4). Fracture corridors, with fracture densities of 5-10 fracture per meter, are common in the portions of the stratigraphy that lack lithologic heterogeneity and are adjacent to normal faults with offset on the cm to 10's m scale, a fracture density ranges from 2-5 fractures per meter, is observed within the siltstone and mudstone interbeds (Figure 4-2, 4-4). At the control outcrop locality, fractures are regularly spaced, terminate at bed boundaries, and have lower fracture densities of 0.3 fractures per meter (Table D-1).

Adjacent to the fracture margins the host rock is bleached from the primary red to yellow/tan (Figure 4-4). This structural diagenesis is caused by exposure to reducing subsurface fluids in zones of enhanced permeability related to fractures. ICPM90A note a loss of up to 30 wt. % Fe between bleached and unbleached samples (Table D-1). The bleached zones range from mm- to m- in width and are generally symmetric zones adjacent to the fractures (Figure 4-4). Whole-rock XRD analysis shows no change in bulk mineralogy between bleached fracture margins and protolith samples (Table D-2).
Figure 4-4. Outcrop photos of fractures and bleached fracture margins within the Organ Rock Formation at locality 1. A) bedding plane view of splay features associated with fracture tips B) outcrop photo of landscape-scale alteration of lower Organ Rock Formation. Note light-colored sandstone is eolian marker bed (~80 m (262 ft) up-section from Cedar Mesa Sandstone). C) cross-sectional view with example of vertical fracture corridor within the Organ Rock Formation.
Optical petrography also shows a change in color is due to the removal of iron oxide (hematite) grain coatings in reflected light (Figure 4-5A). Antitaxial veins are mineralized with hematite and/or calcite, detrital quartz grains adjacent to fracture margins contain vein parallel microfractures with little to no rotation of quartz fragments (Figure 4-5B, Table D-2).

Fracture termination, deflection and/or bifurcation at lithologic interfaces are more common in the thinly interbedded fine siltstone and mudstone facies. Shear fractures observed on bedding plane surfaces have a mean intersection angle of 31° and these fractures curve into the larger continuous NW oriented fractures. At the map-scale horsetail and splay fault terminations are observed at fault tips (Figures 4-1, 4-3A, Table D-1). The mean fracture strike orientation (309°) parallels the joint set of the underlying Cedar Mesa Sandstone (Figure 4-6). The consistent orientation and occurrence of altered fractures within the Organ Rock Formation and the normal faults suggests a linked deformation history. Air permeability and compressive rock strength measurements are limited to the indurated siltstone and very fine sandstones that occur as interbeds throughout the section. All other lithologies are below detection limits of the field equipment.

Quartz-rich Limestone and Shale - Carmel Formation

The stratigraphic section of the Carmel Formation in the San Rafael Swell is gently dipping (9° ± 2° W) and is dominated by argillaceous pelloidal micrite to packstone interbedded with clay-rich shale, and a prominent gypsiferous sandstone
bed. All units except the gypsum marker bed contain mineralized veins (Figure 4-2, Table D-2). Near vertical syntaxial calcite veins indicate that open-mode fractures formed and were the loci of fluid flow within the limestone facies (Figure 4-7); (see Chapter 2).

Fracture density, permeability, and compressive strength vary with changes in lithology and bed thickness, and overall the permeability decreases up-section while bed thickness and compressive strength increases in the packstone and quartz-rich pelloidal micrite layers (Table D-2, Figure 4-8) (Petrie et al., 2012). Calcite veins are found in the resistant lithologies (limestone) whereas fractures in the fine-grained lithologies are characterized by the presence of limonite mineralization along fracture margins, calcite and/or gypsum mineralization in the center (Figures 4-7, 4-8). Veins are observed throughout the Carmel Formation at this and other localities, including core retrieved adjacent to the Big Hole fault, see Shipton et al. (2002) for location information, and in outcrop localities on the steeply dipping eastern side of the San Rafael Swell (Barton, 2011).

The basal portion of the stratigraphic section in the Carmel Formation consist of thinly interbedded argillaceous limestone and shale and is characterized by fracture morphologies that bifurcate, arrest, or are deflected across lithologic boundaries (Figure 4-7). The mean fracture strike orientation is NNE (37°) and is consistent with the orientation of fault deformation bands in the underlying Navajo Sandstone (Figure 4-6), whereas the large-scale joint set in the Navajo Sandstone has a mean strike orientation to the NNW (327°).
Figure 4-5. Organ Rock Formation photomicrographs, A) comparison of bleached and unbleached fracture adjacent samples under reflected light, B) hematite vein with microfractures within quartz grains.
Figure 4-6. Poles to planes of fractures plotted on lower hemisphere stereonet, with Kamb contours, C.I. = 1. Half-circle rose diagrams of fracture strikes with mean orientation as arrow plotted. Estimated orientation of $S_h$ of local and regional tectonic features included for comparison.
Figure 4-7. Carmel Formation vertical outcrop photo showing veins within the quartz rich peloidal micrite and limonite alteration associated with fractures in shale interbed (inset photo). Inset sketch of mineralized extensional-shear fracture.

Structural lineaments within the San Rafael Swell trend NNE and WNW, the hinge-line of the anticline trends NNE (Figure 4-1). Modern maximum horizontal principal stress orientations vary along gently dipping western limb of the San Rafael Swell from NNW within the boundaries of the Colorado Plateau and shifts to NNE along to the Basin and Range/Colorado Plateau transition zone, this transition can be observed in both borehole data of the World Stress Map and fault orientations (Hintze et al., 2000; Heidbach et al., 2008; Janecke unpublished, 2011).
The similarity in structural orientations between fault deformation bands in the underlying Navajo Sandstone and veins within the Carmel Formation suggest their development occurred under the same stress orientations, while the dominant joint set orientation to NNW is consistent with modern stress orientations and likely represents jointing due to uplift/exhumation.

Figure 4-8. Photomicrographs of fractures within the Co-op Creek member of the Carmel Formation. Both photomicrographs show brittle character of the vertical calcite veins. A) Calcite veins cutting quartz rich pelloidal micrite. B) Mechanically twinned calcite vein cutting fossils in bioclastic wackestone.
Siltstone and Mudstone - Earthy Member of the Entrada Formation

This outcrop of the Earthy Member of the Entrada Sandstone in nearly flat lying (dip 5° ± 2°N) and exposes a reservoir-seal pair, where the interbedded fractured siltstone and mudstones of the caprock seal overlie a bleached intraformational reservoir of medium to fine-grained sandstone (Figure 4-9). At this locality the Earthy Member of the Entrada Sandstone is transected by near vertical shear fractures with bleached margins surrounding the veins and fractures. In addition to the bleached margins, there are several active CO₂ springs and several generations of travertine deposits that indicate a history of CO₂ charged spring activity (Dockrill and Shipton, 2010). The active CO₂ springs are on trend with the dominant north-striking fracture set, orthogonal to the fault trace.

Three fracture orientations are observed within the Earthy Member; a minor fault-parallel (296°) striking set, which is cut by a younger landscape-scale north striking (351°) fracture corridors (Figure 4-6, 4-9). The fracture corridors have fracture densities of 10-17 fractures per meter and their strike orientation parallels the axis of the Green River Anticline and is associated with low amplitude anticlines and synclines oriented orthogonal to the Ten Mile Fault (Ogata et al., 2012). The third and youngest fracture set is composed of near horizontal aragonite veins (Figure 4-9C). These veins tip out laterally and generally parallel bedding within the Earthy Member and the overlying Curtis Formation. The steeply dipping north oriented fractures are cut by horizontal aragonite-filled veins that are associated with the modern dissolved CO₂ + water fluid system active in the area (Pearce et al.,
At this locality we observe a consistent relationship between permeability and rock strength across fracture corridors, with permeability increasing and rock strength decreasing across the north striking joint sets (Figure 4-10). Permeability varies between the bleached reservoir (4.935x10^{-15} to 1.184x10^{-13} m^2 / 0.005 to 0.12 Darcy) and unbleached overlying rocks (9.869x10^{-16} to 3.948x10^{-14} m^2 / 0.001 to 0.04 Darcy), with higher permeability values observed within the bleached reservoir (Figure 4-10). Individual fractures are sometimes associated with deformation bands in the underlying thickly-bedded bleached sandstone.

Extensional and shear fractures are surrounded by alteration halos and show evidence for several episodes of fracture reactivation and mineralization. Mineralization within the fractures includes an initial episode of quartz fill, followed by reactivation and mineralization of the joint sets by gypsum and hematite (Figure 4-11). Optical petrography shows a decrease in hematite cement and grain coating between the bleached fracture margins and unbleached Earthy Member host rock. Elemental analysis confirms this with a decrease of 26 wt. % Fe between the unaltered protolith and adjacent bleached fracture host rock (Figure 4-11, Table D-2).
Figure 4-9. Outcrop photos of Earthy Member fracture diagenesis. A) cross-sectional view of vertical fractures cutting upper section of the Earthy Member, B) cross-sectional view at intraformational reservoir seal contact with bleached fractures cutting both the horizontal reservoir bleaching contact and fluid alterations adjacent to vertical fracture margins. C) cross-sectional view of youngest fracture set, horizontal aragonite and calcite veins cut the vertical fracture set.
Figure 4-10. Schmidt hammer compressive rock strength and permeability transects taken across fracture corridors, similar to that shown in Figure 4-9B. There is a consistent relationships between increasing permeability and decrease in rock strength across fracture corridor.
Figure 4-11. Earthy Member photomicrographs. (A) reflected light comparison between unbleached, top, and bleached, bottom. (B) fracture fill showing three phases of fracture mineralization quartz, gypsum, and hematite. Q-quartz, G-gypsum, H-hematite, P-pore space Calcareous Shale - Cretaceous Mancos Shale - Tununk Member
Calcareous Shale - Tununk Member of the Mancos Shale

The Tununk Member of the Mancos Shale is gently dipping (11° ± 4°N) and is cut by near vertical and bed-parallel mineralized fractures. The dominant fracture strike orientation (105°) in the Tununk Member parallels that of the adjacent Little Grand Wash fault zone (Figure 4-6). Discontinuities are dominated by through-going altered fractures or those that terminate into other discontinuities. The fractures are mineralized extensional-shear fractures (Figure 4-12); calcite veins are common with limonite occurring along the fracture margins (Figure 4-12). No field permeability or compressive strength data were acquired at this outcrop as it is below the detection limits of both the field instruments. Petrographic analysis in thin-section and XRD analyses show a mineralogic change at fractures with loss of the platy and clay minerals (mica, illite, kaolinite) and the appearance of calcite. Optical petrography also shows veins of mechanically-twinned calcite, shear fabrics, and hematite lining fracture margins (Figure 4-13, Table D-1). Fracture spacing at the study locality is 0.71 fractures per meter and fractures are easily identified by the limonite margins and associated color change. At the control outcrop locality, fractures are evenly spaced, with 1 fractures occurring every meter.
Figure 4-12. Tununk Member cross-section of outcrop exposure in the fault damage zone of the Little Grand Wash Fault.
Figure 4-13. Photomicrographs from the Tununk Member. (A) S-C fabric of shear zone enlarged and false color to highlight the S-C fabric. (B) Hematite vein interactions viewed in reflected and cross-polarized light. (C) Vein mineralization includes both of crack seal textures of calcite and hematite.
Burial and Stress History Models

The field data presented here show that hydraulic extension fractures, extensional-shear fractures, and extensional mesh fracture networks transect at minimum 17 m (55.8 ft) of the sealing lithologies and serve as conduits to fluid flow. We consider a uniaxial strain model and use changes in stress magnitude through time to understand the potential for formation of these brittle structures within varied lithologies under lithostatic loads (Figure 4-3). Because we observe veins and evidence for subsurface fluid migration we model the effect of increased pore pressure on mechanical rock failure and mode. Depending on rock mechanical properties and the magnitude of differential stress when $\sigma_3$ is less than zero, extensional-shear failure or hydraulic extension can occur (Sibson, 1985, 2003). Extensional-shear fractures open perpendicular to $\sigma_1$ when $\sigma_3$ is tensile or when $\sigma_3$ is compressive but fluid pressure is sufficient to make the effective stress ($\sigma_3 - P_f$) tensile (Laubach, 1988; Sibson, 2003).

Results from the indirect tensile strength test show clear variations in rock strength due to lithologic changes in mineralogy and grain size (Figure 4-14, Tables D-1, D-3). The calcareous shale of the Tununk Member and the micritic limestone of the Windsor Member of the Carmel Formation have the highest tensile strength (10-12 Mpa / 1450-1740 psi) whereas the siliciclastic dominated lithologies, the Organ Rock Formation and Earthy Member have the lowest (2-3 MPa / 290-435 psi); the quartz-rich peloidal micritic limestone of the Co-op Creek Limestone Member of the
Carmel Formation has an intermediate tensile strength with a median value of T=5 MPa (725 psi) (Figure 4-14).

We use the median value of tensile strength obtained for each lithology, in a Mohr-Coulomb-Griffith failure analysis to understand the change in pore-fluid pressure required to induce failure through time. This method allows us to examine the conditions for which creation and/or reactivation of brittle structures may occur.

Figure 4-14. Results of indirect tensile strength tests. Individual results for each unit shown; dashed lines represent median values used in failure analysis.
We use the pore-fluid factor to compare the fluid pressure required for failure throughout the modeled stress histories. The pore-fluid factor, $\lambda_v$, is the fluid-pressure ($P_f$) in relation to the overburden pressure ($\sigma_1=\sigma_v$) calculated as (Ingram and Urai, 1999; Sibson, 2003);

$$\lambda_v = \frac{P_f}{\sigma_v}$$

(5)
Pore-fluid factor values approach but do not exceed 1 in extensional tectonic regimes and $\lambda_v$ values greater than 1 are only achieved at compressive stress regimes with low differential stress magnitudes (Sibson, 2003). We calculate the pore-fluid factor through time for each seal analog modeled as the relationship between the $P_f$ required for failure relative to the estimated maximum stress defined by the burial history curve.

The calculated values of $\lambda_v$ show that it approaches 1 at shallow depths as expected for failure in an extensional regime. We note that the changes in mechanical properties associated with each lithology play an important role in the $\lambda_v$ value and failure mode. For intact rocks the quartz-dominated, coarser grained facies with lowest tensile strengths are predicted to fail in extensional shear when $\lambda_v$ at depths greater than 1.5 km (0.932 miles) (Figure 4-15, Table D-3). At depth values less than 1.5 km (0.932 miles), $\lambda_v$ is $\geq0.65$ and mechanical failure occurs in hydraulic extension for all lithologies and modeled scenarios. At burial depths greater than 1.5 km, the quartz-rich siltstone facies fail in extensional shear while the calcareous rock types continue to fail in hydraulic extension. Burial
depths greater than 3.5 km are required for the calcareous rock types to fail in extensional shear (i.e. high differential stress conditions) (Figure 4-15).

The calculated pore-fluid factors for failure in the cohesionless case range from 0.38-0.64 with a median of 0.34 and reactivation of optimally oriented existing structural discontinuities results in extensional-shear fractures, hydraulic extension fractures do not occur (Sibson, 1981, 2003). The \( \lambda_v \) values in the cohesionless scenarios are very near hydrostatic pressure, \( \lambda_v \) for the hydrostat pore fluid pressure is 0.385.

![Figure 4-15. Pore-fluid factor values for each lithology derived from burial history and Mohr-Coulomb failure analysis, (burial history and failure analysis for each locality can be found in Appendix Table D-3). Pore-fluid factor (\( \lambda_v \)) values less than 0.65 result in extensional-shear failure; while values greater than 0.65 result in hydraulic extension fractures, shear failure occurs at \( \lambda_v \) values \( \leq 0.385 \).]
Discussion

As shown by outcrop and petrographic analysis, each seal lithology examined exhibits features of mechanical failure and fluid flow in the subsurface. Mineralized veins provide evidence for past seal failure and the variability in morphology suggests that the inherent lithologic heterogeneity of the seal imparts varied mechanical properties over small scales (10 cm to 1 m / 3.9 in to 3.28 ft) as also recognized by (Larsen et al., 2010). We observe hydraulic extension fractures, extensional-shear fractures and extensional mesh fracture networks in outcrop and show that these fracture networks occur across varied caprock seal lithologies characterized by different rock strengths. Changing mechanical properties across lithologic interfaces within the seal results in penetration, deflection, or termination of fractures at lithologic interfaces (Figures 4-4, 4-7, 4-8, 4-9,4-12,4-13) (Larsen et al., 2010).

We observe fracture corridors where the stratigraphic section lacks lithologic heterogeneity and are adjacent to normal faults (Figures 4-2, 4-4, 4-9). Fracture strike orientations within each outcrop locality are consistent with the associated map-scale structures (Figure 4-1, 4-6, Table D-1). Fracture strike orientations within the Organ Rock Formation are consistent with low-offset (<30m / 98.4 ft) long strike length normal faults (>25 km / 15.5 miles); fracture strike orientations within Carmel Formation are consistent with fault deformation bands in the underlying Navajo Sandstone as well as the orientation of the axial plane of
the San Rafael Swell; the dominant fracture strike orientation data within the Earthy Member is parallel to the axial plane of the Green River anticline and orthogonal to the Ten Mile fault; and fracture strike orientation data from the Tununk Member is consistent with the orientation of the Little Grand Wash fault which has up to 210 m (689 ft) of offset (McKnight, 1940; Shipton et al., 2005). The consistent orientation data between map-scale structural features and the observed seal bypass fracture systems as well as the cross-cutting relationships and punctuated events of vein mineralization suggest a history of reactivation under regional tectonic stress.

In a critically stressed brittle crust (Barton et al., 1995), episodically varying fluid pressure through time can result in crack-seal propagation (Ramsay, 1980; Engelder, 1985), with optimally oriented faults and pre-existing fractures prone to failure prior to the formation of new hydraulically induced fractures (Laubach, 1988; Capuano, 1993; Gale et al., 2007). The presence or absence of pre-existing brittle structures, the tectonic regime, and the magnitude of differential stress controls the maximum overpressure intact rock can withstand (Sibson, 2003).

Combining the stress history analysis and rock properties specific to each sealing lithology we show that cohesionless or critically stressed pre-existing structural discontinuities will fail under relatively small changes in pore fluid pressure above hydrostatic pressure (Table 4-3). Pre-existing fractures are considered here to be cohesionless and as such the maximum overpressure they can withstand will be lower than intact rock and their expected mode of failure is
extensional-shear not as hydraulic extension fractures. For the cohesionless case, we predict that a very small increase (0.1 to 1.5 MPa / 14.5 to 218 psi) in pore fluid pressure results in failure in all cases (Barton et al., 1995; Zoback and Townend, 2001) (Figure 4-15, Table D-3). For intact rock, pore fluid pressures required for failure range from 6 to 25 MPa (870 to 3626 psi) and are dependent on burial depth and mechanical rock properties.

In the case of intact rock we shows that the pore-fluid factor, $\lambda_w$, decreases with depth and approaches unity at shallow crustal depths which is consistent with a normal fault stress regime (Sibson, 2003). These uniaxial strain model scenarios highlight the dependence of the maximum sustainable overpressure upon mechanical rock properties and the magnitude of the differential stress. When $\lambda_w$ is less than 0.65 extensional-shear fractures occur within the siliciclastic rock types of the Organ Rock Formation and the Earthy Member (Figure 4-15) (Table 4-3). Rock strength parameters are greater in the calcareous lithologies of the Carmel Formation and the Tununk Member and due to their mechanical properties require higher differential stress conditions to achieve extensional-shear failure (Figure 4-15, Table D-3).

The analyses presented here highlight the importance of understanding the mechanical properties specific to each caprock seal and the rock type distribution within the seal for evaluation of seal integrity or production of fluids from self-sourced reservoirs. We show that all fine-grained lithologies do not behave in the same manner mechanically. In the presence of pre-existing, optimally oriented
cohesionless planes of weakness, mechanical failure will occur with small increases in fluid pressure above the hydrostatic pressure. Additionally these zones of weakness, if not optimally oriented are likely to add complexity to the subsurface fracture network. During fracture growth, hydraulic fractures interact with pre-existing sealed fractures resulting in widespread fracture distributions (Gale et al., 2007). Knowledge of the tectonic history and burial history can provide controls on expected pre-existing discontinuity orientations.

Conclusions

Mode of mechanical failure, hydraulic extension or extensional shear are controlled by the interplay between material properties inherited during the depositional and tectonic history. We show that for intact rock scenarios mechanical properties are an important control on failure mode, i.e. the occurrence of hydraulic extension or extensional shear. Hydraulic extension is predicted to occur preferentially in rock with large tensile and compressive strengths, in this case calcareous lithologies while in rocks with moderate tensile and compressive strengths shear fracture are dominant, quartz rich lithologies. The data presented here shows the importance of understanding the stratigraphic distribution of varied lithologies within the formation of interest, allowing for changes in failure mode expected under specific conditions in the subsurface.
In our modeled stress history analysis we consider pre-existing fractures to be cohesionless, reactivation of pre-existing hydraulic extension fractures or extensional-shear fractures (Table D-1). Optimally oriented fracture sets will fail, particularly at fluid pressures very near hydrostatic (Table D-3). Although the pore-fluid factor decreases with increasing depth for all intact rock scenarios, differences in mechanical properties due to lithologic changes inhibit extensional shear failure by increased pore fluid pressure alone. Both calcareous facies, the Tununk Member and Co-op Creek Limestone Member reach similar maximum burial depths as the quartz-rich facies of the Organ Rock Formation and Earthy Member, however due to the rock properties failure occurs in extensional-shear regime (Figure 4-15, Table D-3).

The estimates for failure based on rock mechanics and burial history derived stress are minimum case observations; regional tectonic stress magnitudes were not modeled. In outcrop the observed deformation features are likely reactivated, optimally oriented crack-seal fractures that provide zones of weakness, and/or localized stress for normal faulting or the normal faults were optimally aligned and reactivated due to increased pore fluid pressure.

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CHAPTER 5

Modeling the Role of Mechanical Interfaces to Fracture Propagation and Morphology

- Geomechanical Models Derived from Outcrop Analysis¹

Abstract

In the brittle crust the rock strength is a function of physical properties and their interactions with overburden pressure, fluid saturation, and tectonic loading. Mechanical failure in the brittle crust may occur by slip across faults or by propagation of fracture networks. Both of these failure modes are important for formation of fluid-flow pathways in low-permeability rocks. In order to document the effects changing mechanical properties have on the distribution of fractures in the subsurface we use previously identified mechano-stratigraphic units in outcrop and elastic moduli derived from borehole geophysical logs to populate three-dimensional finite element models.

We model how variations in stratigraphic thickness and changes in elastic moduli impact the distribution of strain within sedimentary sequences and across interfaces. We use finite element analysis to model plane-strain conditions where vertical and horizontal stress magnitudes are fixed while varying unit thickness and elastic moduli in accord with elastic moduli derived from offset geophysical wireline

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Co-authors listed as: ES Petrie and JP Evans
Department of Geology, Utah State University 4505 Old Main Hill, Logan, UT 84322-4505
logs. This model is of a relatively small area (10m thick) and involves stratigraphic changes (1m) that can not be detected using reflection seismic surveys. In outcrop and core sedimentary interfaces mark the boundary between mechanical units and fractures deflect, propagate across, or arrest at these boundaries.

We compare the modeled strain distribution results to the outcrop fracture distribution. The models serve as a proxy for changes in expected fracture distributions within and across model horizons and are compared to outcrop fracture distribution. The results show that variability in strain distribution can be used to predict the natural deformation response manifested as more fractures in high strain regions and fewer fractures in low strain regions.

A simple 3-layer model was developed to mimic the mechano-stratigraphic units identified in outcrop. The 3-layer model misses the thin-bed effects associated with alternating lithologies within the Carmel Formation and does not adequately predict the high variability of fracture density observed in outcrop. In contrast the 5-layer model detects marked changes at material interfaces and is a better match for outcrop distribution of fractures. The field observations and modeling results indicate that the potential for subsurface failure and fluid flow would not be restricted to the low fracture strength units but could occur across the seal.

Because outcrop evidence shows fractures interact with sedimentary interfaces and that seals contain fractures and small faults the modeling results can
be used predict zones of higher strain and therefore higher expected fracture densities or layers that are more prone to fracture formation.

Introduction

The presence of structural discontinuities such as joints, veins, or small-displacement faults within fine-grained lithologies can act as loci for fluid flow over various timescales. Open-mode fractures can propagate under low strain conditions (Olson et al., 2009) and fractures may act as important permeable pathways for subsurface fluid flow (Laubach, 2003).

Impermeable caprocks act as top seals for conventional hydrocarbon resources and are becoming increasingly important as unconventional hydrocarbon reservoirs and as top seals of waste repositories in the subsurface (Herzog, 2001; Warpinski et al., 2009), therefore understanding the mechanical response of fine-grained lithologies to changing stress conditions and the impacts of changing material properties is essential.

In this paper we present results from geomechanical models that encompass the basal 9 meters of the Carmel Formation (caprock seal) and upper 3 meters of the underlying Navajo Sandstone (reservoir). The models and outcrop data presented here evaluate interfaces that are below resolution limits of most seismic reflection data (<10m) but are resolvable with borehole geophysics. These models consider two parameters: changes in elastic properties and unit thicknesses. Variations in
local stress orientations and/or tensile stresses at the fracture tip have been considered previously (Gudmundsson, 2009; Larsen et al., 2010).

The possibility of the formation of new or the reactivation of existing fractures in fine-grained lithologies requires development of accurate geomechanical models that utilize actual mechanical properties of real-world seals. We populate the models with data derived from both surface exposures and nearby drill holes of a reservoir and its overlying seal. This seal was chosen because it shows evidence for mechanical failure and subsurface fluid flow and its mechanical properties were previously derived using geophysical wireline log data (see Chapter 2). In this paper we use a history matching technique where the finite element models (FEM) are created to best match observed and measured data (Schlumberger Oil Field Glossary, 2013).

The dynamic elastic moduli calculated from wireline logs (Young’s modulus and Poisson’s ratio) were used to quantify the isotropic, elastic mechanical properties of the lower Carmel Formation. These data sets are combined with the interpreted mechnano-stratigraphy to evaluate the potential for deformation (manifested as fractures) to exist in the subsurface. We use the geo-mechanical models to evaluate meter-scale changes in mechanical properties and compare the modeled strain distribution to the deformation features observed in outcrop in order to understand the model fit and its predictive abilities with deformation features observed in the outcrop analog.
Geologic History

The Jurassic Carmel Formation is a fine-grained mixed siliciclastic carbonate sedimentary sequence that at the study locality represents deposition in the near-shore marine to sabkha environments (Blakey, 1994; Caputo, 2003; Hintze and Kowallis, 2009). The Carmel Formation is an excellent field analog for the study of a heterolithic seal. The Carmel Formation, at the outcrop locality along I-70, is well exposed on the gently dipping western limb of the San Rafael Swell, Utah (Figure 5-1). The San Rafael Swell is an asymmetric, east-vergent, doubly plunging anticline with a NNE-trending hinge-line (Gilluly, 1929; Bump and Davis, 2003; Davis and Bump, 2009).

The extension direction changes by about 90° in the western third of the San Rafael Swell from a stress field of the Colorado Plateau across most of the structure to one of the Basin and Range on the lower part of its northwest flank. Small normal and normal-oblique faults strike ENE and WNW across most of the Sell and NNE in the western part around the outcrop analog (Figure 5-1) (Kelley and Clinton, 1960; Krantz, 1988; Witkind, 1988; Shipton and Cowie, 2001; Doelling, 2002b; Bump and Davis, 2003; Janecke, personal communication, 2012). Modern maximum principal stress orientation along the gently dipping western limb of the San Rafael Swell NNE (~8°) (Heidbach et al., 2008). Estimated timing of formation of the San Rafael Swell ranges from ~93 Ma to 58 Ma (Fouch et al., 1983; Lawton, 1985; Molenaar and
Cobban, 1991; Guiseppe and Heller, 1998; Shipton and Cowie, 2001), due to the east-west directed compression during the Sevier and Laramide orogenic events.

Figure 5-1. Generalized geologic map showing outcrop location, faults and approximate axial surface of the San Rafael Swell. Map modified from (Hintze et al., 2000; Hintze and Kowallis, 2009; Barton, 2011; Petrie et al., 2013)
Key Field Observations and Interpretations

At the study locality the Carmel Formation dips gently (9 ± 2°W) and overlies the Navajo Sandstone (Figure 5-2). Prior analysis characterized the entire 37 m thick exposure (Figure 5-3). The Co-op Creek and basal portion of the Crystal Creek members of the Carmel Formation are thin- to medium-bedded, quartz-bearing pelloidal micrite, thin- to medium-bedded bioclastic wackestone and calcareous mudstone and shale dominate and are considered in this study to be the primary seal. For this and computational reasons, we produced a finite element model of the lower 9m of the seal and upper 3 m of the reservoir. In outcrop veins cross-cut lithologic boundaries and extend from the reservoir seal interface 5-10m into the overlying Carmel Formation.

Figure 5-2. Outcrop photo and stratigraphy of Carmel Formation and underlying Navajo Sandstone.
Field-derived rock strength and permeability estimates vary stratigraphically within the Carmel Formation (see Chapter 2). Compressive rock strength, determined from Schmidt hammer rebound values, exhibits higher variability in the thinly bedded heterolithic portion of the lower Carmel Formation. The inherent lithologic heterogeneity of this portion of the Carmel Formation imparts varied mechanical properties over small scales (10 cm to 1 m). Calcite veins occur in the resistant lithologies (limestone) whereas fractures in the finer grained lithologies exhibit limonite, calcite, and/or gypsum mineralization.

Outcrop data delineated 5 mechano-stratigraphic units within and exposure of the Carmel Formation, located in central Utah (see Chapter 2). These mechano-stratigraphic units are based on similarities in fracture spacing, bed thickness, Schmidt hammer-derived compressive strength, and air permeability measurements (see Chapter 2).

The mechano-stratigraphic units modeled in this paper include mechano-stratigraphic units 1, 2, and 3. The lowermost mechanical unit, unit 1, (Figure 5-3), is the Navajo Sandstone reservoir, a high permeability, thick-bedded, low fracture density quartz arenite. The Navajo Sandstone has a mean fracture density of two fractures per meter (fracture/m) and contains open joints and fault deformation bands. The overlying Carmel Formation is split into four mechano-stratigraphic units, labeled 2-5 (Figure 5-3).

Unit 2 is composed of interbeds of thin-bedded quartz-rich limestone, siltstone and shale. This unit coarsens upward and becomes more fossiliferous up-
section. Unit 3 is finer grained; more thinly bedded and has a higher fracture density than the underlying unit 2 (Figure 5-3). Veins, step-over fractures, and bifurcated fractures are common in mechano-stratigraphic units 2 and 3, and bed parallel fractures are observed within the shale layers (Figure 5-3). In thin-section mechanical twins are observed within the calcite veins of the Carmel Formation. Their presence of which indicates that fracture opening and mineralization occurred prior to being deformed at some depth. The near vertical calcite veins contain symmetric mineralization patterns indicating that open mode fractures formed and were the loci of fluid flow at depth. Similar vertical calcite veins occur throughout the Carmel Formation at this and other localities in the San Rafael Swell (Barton, 2011).

Fractures have two main strike directions are recorded at this outcrop. The Navajo Sandstone and Carmel Formation share NNW and NNE fracture orientations (Figure 5-4). The dominant fracture strike is NNW, and the majority of fractures within the Carmel Formation, including veins, and fault deformation bands in the Navajo Sandstone share this orientation.

The similar orientation of discontinuities and deformation band trends in the Carmel Formation and Navajo Sandstone suggests a similar history of formation and timing of deformation. The veins observed in the Carmel Formation and the fault deformation bands in the Navajo Sandstone likely represent paleo-stress directions remnant of the Laramide uplift and associated deformation along the western edge of the San Rafael Swell (Anderson and Barnhard, 1986; Davis and Bump, 2009). We
interpret the NNE fracture set in both the Navajo Sandstone and overlying Carmel Formation to reflect the modern maximum principal stress orientation (Heidbach et al., 2008) and are likely related to Cenozoic tectonic deformation.

We note that the shale horizons of the Carmel Formation exhibit a greater dispersion in strike orientations than the limestone and sandstone lithologies (Figure 5-4B). This data dispersion and its association with lithology suggests that each lithology responds differently to stress. In a stratigraphic sequence with high lithologic variability highly variable fracture distributions are likely and will affect fracture patterns and in turn fluid flow pathways at depth. This variability may be due to the development of tensile stress ahead of the fracture tip, rotation of localized stress orientation at lithologic boundaries, and/or variation in elastic properties across interfaces (Larsen et al., 2010). The models presented here evaluate the effect variations in elastic properties may have on strain distributions at and across interfaces.

In the models presented here we evaluate variations in elastic properties across interfaces within the lower 9 meters of the Carmel Formation and uppermost 3 meters of the Navajo Sandstone. We note in the field that the stratigraphic heterogeneity has influenced the variability in fracture pattern, rock strength and permeability throughout this exposure of the Carmel Formation (see Chapter 2). Higher fracture densities, permeability, and compressive rock strength is observed in the thinly bedded heterolithic facies of mechano-stratigraphic units 2 and 3 of the Carmel Formation and continuous fractures are observed from the reservoir as fault
deformation bands and as open mode or shear fractures in the overlying sealing stratigraphy.

Modeling Methods

We combine previously defined mechano-stratigraphic observations with calculated dynamic elastic moduli derived from offset exploration and production boreholes to create geomechanical models that cross the reservoir cap-rock seal interface and that capture the heterogeneity within the lower 9 meters of the Carmel Formation and include the upper 3 m of the Navajo Sandstone. Dynamic values for Young’s modulus, (E), and Poisson’s ratio, (ν), have been derived from wireline well-log analysis using a combination of 20 exploration boreholes and production wells (Figure 5-5). Using these 20 boreholes we estimate dynamic value of elastic moduli within the Carmel Formation by building on specific empirical relationships between lithology to compressional and shear velocity (Pickett, 1963; Castanga et al., 1985; McCann and Entwilse, 1992). For details on methodology see Chapter 2. In the models presented here we use dynamic elastic moduli derived from Celsius Fed 8-1 well shown in Figure 5-5, ~12 km from the outcrop location.
Figure 5-3. Stratigraphic column of the Carmel Formation including rose diagrams of structural discontinuity data from scan lines and field photographs of cm-scale discontinuities observed in outcrop. Previously determined mechano-stratigraphic units from see Chapter 2 are highlighted and labeled 1-5. These mechano-stratigraphic units were determined using similarities in bed thickness, fracture distribution, air permeability, and Schmidt Hammer rebound values for rock compressive strength.
Figure 5-4. A) Rose diagram of strike orientation data showing the two fracture sets and common orientations observed between the Navajo Sandstone and Carmel Formation. B) Stereographic projection of poles to planes of fractures within the limestone and shale facies of the Carmel Formation.
Figure 5-5. Borehole data used in analysis of elastic moduli. Data from Celsius Fed 8-1 was used in the models presented here. Map modified from (Hintze et al., 2000; Doelling, 2002a; Chapter 2)

Physical properties for each model are based on field observations and delineation of mechanical stratigraphic units as well as variability in Young's modulus (E) and Poisson's ratio (ν) derived from geophysical wireline logs. We compare two models: Model I is a three-layer model based on field observations of mechanical stratigraphy alone and Model II which is based on a combination of the
outcrop defined mechanical stratigraphy and observed variability in calculated elastic moduli from geophysical logs (Figure 5-6). The physical properties (layer thickness, $E$, and $v$) vary between the model scenarios as listed in Table 5-1.

Table 5-1. Model parameters.

<table>
<thead>
<tr>
<th>Layer Properties</th>
<th>Unit thickness (m)</th>
<th>Poisson's Ratio</th>
<th>Young's Modulus (GPa)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Model I</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>4.0</td>
<td>0.29</td>
<td>25</td>
</tr>
<tr>
<td></td>
<td>5.0</td>
<td>0.29</td>
<td>23</td>
</tr>
<tr>
<td></td>
<td>3.0</td>
<td>0.29</td>
<td>18</td>
</tr>
<tr>
<td>Model II</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>4.0</td>
<td>0.29</td>
<td>33.7</td>
</tr>
<tr>
<td></td>
<td>1.0</td>
<td>0.26</td>
<td>17.2</td>
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<td>0.30</td>
<td>26</td>
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<td>0.28</td>
<td>17.2</td>
</tr>
<tr>
<td></td>
<td>3.0</td>
<td>0.21</td>
<td>18.5</td>
</tr>
</tbody>
</table>

Using the free student version of AbaqusFEA® (DassaultSystemes, 2011), a finite element analysis (FEA) software program, we create geomechanical models for elastic deformable, layered-solids. We examine a model domain of length 5 m, total height 12 m, and depth of 5 m (Figure 5-7). In the student version of AbaqusFEA® the model domain is limited to analysis of 1000 nodes and following engineering convention the numerical strain values are negative when shortening and positive when in extension. The model solves the constituitive equations for elastic solids at the nodes subjected to the boundary condition loads, and the strain values can be determined anywhere in the block.
Figure 5-6. Model layering schematic depicting the observations in outcrop and wireline logs used to populate the model domain with mechanical properties and unit thickness. A) Model I is a simple 3-layer model based on mechano-stratigraphic units identified by outcrop analysis. B) Model II is based on observed shifts in lithology from Gamma Ray logs, and variations elastic properties.
Figure 5-7. Cut diagram of undeformed model shape showing mesh discretization and node based observation points taken near the center of each model. In the model space x and y are horizontal directions and z is the vertical dimension.

To estimate the magnitude of applied stress in the multi-layered scenarios we consider uniaxial strain conditions and calculate the magnitude of vertical ($S_v$) and horizontal ($S_h$) stress experienced in the subsurface at maximum burial depth. Maximum burial depth was determine by creating a burial history curve using OSXBackStrip (Cardozo, 2010) and is based on the compiled stratigraphic section.
for the San Rafael Swell (Hinzte and Kowalis, 2009) the curve is uncorrected for compaction. We assume an Andersonian tectonic stress orientation for a normal fault regime (Sv=σ1); where

\[ S_v = pz \]  
\[ S_H = S_h = \sigma_2 = \sigma_3 = \left( \frac{\nu}{1-\nu} \right) (S_v - P_p) \]

(Anderson, 1951; Eaton, 1969; Engelder, 1993). The magnitude of Sv is based on the lithostatic load (equation 1) at an estimated maximum burial depth of 4.4 km.

We model two scenarios; each with bed thickness variations that are too small to be detected by high resolution reflection seismic data, but in outcrop show variability in fracture distribution and whose geophysical wireline data show clear variations in elastic moduli (Figure 5-6C & 5-7). Model I has a layer thicknesses based on outcrop mechano-stratigraphic divisions. Estimates of elastic moduli for model I are averaged from the wireline logs, that is the upper 3 meters of the Navajo Sandstone reservoir, and two sections within the Carmel Formation of 4 meters and 5 meters each (Figure 5-6 & Table 5-1). In contrast model II uses the average elastic moduli in the upper 3 meters of the Navajo Sandstone and then variability observed in Gamma Ray lithology, Young’s modulus and Poisson’s ratio within the Carmel Formation. Model layers where chosen using changes in Gamma Ray lithology that where associated with shifts in Young’s modulus ≥10 GPa and over a wireline log thickness greater than 61 cm (2 ft) and a change in Poisson’s ratio (Figure 5-6C & Table 5-1). This variability within Model II adds two layers to the middle of the model, and correlates to lithologic changes within mechano-stratigraphic unit 2 & 3.
This results in a more sophisticated model that explicitly breaks out the fine grained intervals in the Co-Op Creek and lower Crystal Creek Members.

In each FEM scenario the grid blocks are defined as 1m cubes, and strain values are computed at each node point where node points are defined at the intersection points of each grid block (Figure 5-7). Each modeling scenario has a fixed base and is under vertical and horizontal loads defined by the estimated $S_v$ and $S_h$ values. The vertical load estimate $S_v$ is 113 MPa and $S_h$ is estimated to be 66.5 MPa where Poisson’s ratio, $\nu = 0.25$ (Equation 2 & Figure 5-8). The FEA was conducted in a single step applying the vertical and horizontal stresses as uniform pressure to all sides except the base. To replicate plane strain, deformation was allowed to occur in all directions except the x direction, oriented normal to the vertical load.

Strain values from the resultant models are displayed as the maximum principal strain taken at the vertical observation points near the center of each deformed model space (Figure 5-7, 5-9 & 5-10). We evaluate the strain response of each layer with regard to the elastic properties to understand the interplay between elastic moduli and resultant deformation. We then compare the strain distribution results from each model to our observed outcrop fracture distributions, to understand the importance of heterogeneity and the detail required to capture strain distributions that are representative of the observed fracture distribution. By comparing the strain distribution to the observed fracture distribution we are employing a history matching technique, borrowed from reservoir engineering,
where the building of numerical models is done to account for observed and measured data (Schlumberger Oil Field Glossary, 2013). In this case we use the FEM results to evaluate the detail required to predict the presence of fracture or potential for inducing fractures. We assume layers with higher fracture densities experienced more strain thus allowing for comparison to the model results.

Figure 5-8. Model configuration for the FEM scenarios. A) Model I - three-layer model and B) Model II – five-layer model. Both models have a distribution of material properties based on layer thickness, a fixed base, a vertical stress equal to $S_v$, and horizontal stress on all vertical margins. No deformation occurs in the x-axis direction, normal to $S_v$, representing a plane strain scenario.
Results and Discussion

The two model configurations predict higher strain values within the caprock seal than the basal reservoir and strain values in general increase upward from the model base. Overall, both model scenarios show localized strain distributions with marked transitions across the reservoir seal interface (Figures 5-9 & 5-10).

Model I depicts a 3-layer sequence with values of $E$ and $\nu$ are averages derived from wireline logs taken over the unit thickness defined by outcrop mechanical stratigraphy alone (Figures 5-6 & 5-8). The resulting strain distributions within the layers show lowest strain magnitude within the reservoir, a marked increases at boundary between the reservoir and overlying seal, followed by a strain decrease upward through the remaining layers. The similarities between elastic moduli, especially Poisson’s ratio, within the seal prevent the model results from depicting meaningful differences in strain across interfaces within the caprock seal. The scale at which the values were averaged across the geophysical wireline logs results in loss of heterogeneity and when compared to observed outcrop is not representative of the variability observed in fracture density. However, the higher strain values within the cap-rock seal relative to that of the underlying reservoir suggest that the seal would be prone to fracture or has been fractured in the past.

In contrast Model II depicts 5-layers in which mechanical properties are based on shifts in dynamic elastic moduli values and lithologic variations observed in the wireline logs (Figure 5-6).
Figure 5-9. Numerical modeling results for plane strain Model I. Vertical strain profile values derived from observation points taken near the center of the model, shown by red line on block diagram. Fracture density from outcrop scanline data shown at far right.
Heterogeneity of this model is increased by adding two low E and moderate v layers which characterize the shale rich units of the lower Carmel and in the model are isolated by stiffer, more incompressible layers (higher E and v values) (Figure 5-7).

When compared the resultant vertical strain profiles from Models I & II highlight the importance of using accurate estimates of both Poisson's ratio and Young's modulus in geomechanical modeling scenarios. Model II shows that variation in values of Poisson's ratio in combination with Young's modulus play an important role in strain distribution.

In Model II both units 2 and 4 have v values larger than that of the underlying reservoir but lower than adjacent beds within the seal and smaller E values relative to the reservoir and surrounding seal. The heterogeneity in mechanical properties and the interaction between changing values of Poisson's ratio and Young's modulus results in: 1) a decrease in positive strain values at the reservoir seal, the unit 1 / unit 2 interface, 2) an increase in strain between unit 2 and unit 3 which corresponds to an increase in observed fracture density in outcrop, and 3) a marked increase in strain values within unit 4 (Figure 5-10). These results suggest that fractures would not necessarily tip out within the weak shale layers. Unit 5 shows an overall decrease in strain up-section, which is consistent with observed outcrop fracture densities. The overall decrease in strain at the top of the model occurs within the high v and E layer may be due to its more incompressible nature and layer thickness (Gross, 1993; Bai and Pollard, 2000).
In Model II abrupt changes in strain values are observed across all mechanical unit interfaces. Using the nature of strain distribution in these models as a proxy for fracture distribution we expect the regions of higher or increasing strain values across interfaces to indicate layers that are more likely to have failed or will fail in future. Model II results capture more variability in strain distribution as well as larger shifts in strain values at and across interface boundaries than Model I. The additional layers in this model provide an overall better match to the observed fracture distributions in outcrop.

Although the changes in mechanical properties are not large across the layers of the models presented here, the resultant strain distributions suggest 1) strain magnitudes change at interface boundaries 2) thin weak layers do not prevent fracturing or inhibit propagation of fractures across interface boundaries (Rijken and Cooke, 2001; Larsen et al., 2010), 3) fractures can be widespread within a heterolithic seal, and 4) the interplay between elastic moduli effect strain distribution, requiring realistic values of these properties be applied to model scenarios. The field observations and modeling results indicate that the potential for subsurface failure and fluid flow would not be restricted to the low fracture strength units but the could occur across the seal.
Figure 5-10. Numerical modeling results for plane strain Model II. Vertical strain profile values derived from observation points taken near the center of the model, shown by red line on block diagram. Fracture density from outcrop scanline data shown at far right.
Conclusions

Studying the occurrence of, and changes in, outcrop fracture patterns and scaling these up for field-scale (km-scale) modeling is difficult due to the lack of direct correlation between outcrop observations and subsurface data. We bridge this correlation gap by first evaluating the meso-scale (mm to m) variability observed in outcrop and incorporate these observations into geomechanical models in order to identify the controlling properties in fracture propagation and morphology. Integration of field (analog) and subsurface datasets for appropriate modeling of the geomechanical response of heterolithic, fine-grained, low-permeability rocks is key in producing accurate model results. We show here that the more detailed model scenario, conditioned on outcrop or subsurface constraints for the fundamental elastic moduli, is a more accurate predictor of subsurface strain distribution and expected deformation. The widespread distributions of strain in both models suggest that fractures would propagate across mechanical boundaries; this result is supported by field observations.

The two finite element models presented here reveal that strain values vary within the different horizons, suggesting that the interaction between the elastic moduli play an important role in the distribution of strain in the subsurface. We also observe variations in strain magnitudes across locked mechanical interfaces, suggesting that an elastic mismatch between layers can result in significant changes
to strain distribution in the subsurface. The results show that variability in strain distribution can be used to predict the natural deformation response manifested as more fractures in high strain regions.

Although neither model is able to replicate some of the thin bed effects, of high fracture density associated with beds less than 0.5 m thick, our use of both mechanical stratigraphy and correlation to wireline-derived elastic moduli allowed for better overall prediction of strain distribution in the subsurface. Elastic mismatch across interface leads to strain differential and the model results highlight the importance of rock properties to understanding fracture distribution in the subsurface.

Our characterization of variability in rock strength and the associated changes in subsurface strain distribution is especially important for modeling the response of low-permeability rocks to increased pore pressure, and is applicable to multiple geo-engineering scenarios such as exploitation of natural resources, waste disposal, and management of fluids in the subsurface. Our analyses provide analog fracture data for heterolithic seals and a dataset for better understanding the importance of changing mechanical properties in low permeability rocks.

References Cited


Gilluly, J., 1929, Geology and oil and gas prospects of part of the San Rafael Swell, Utah: U.S. Geological Survey Bulletin 806-C.


CHAPTER 6

CONCLUSIONS

The key conclusions from the work presented here are:

1. All caprock seal analogues analyzed in this study show a dominance of small fracture spacing (<0.5 fractures per meter) and evidence for failure and fluid flow in the subsurface, with continuous mineralized fractures occurring well above the reservoir/seal interface.

2. The presence of these fracture networks increases the volume of rock matrix that is in contact with subsurface fluids; these fracture networks create preferential pathways for fluid flow and alter the mechanical properties of the seal.

3. Stratigraphic variability and tectonic setting influences fracture density with observed fracture densities higher adjacent to faults, in thin beds (<0.5 m), and in indurated layers.

4. Clastic rocks contain lithologic heterogeneities due to changes in mineralogy and grain size, and these changes result in diverse mechanical properties that lead to variable connectivity of fractures and variable fracture densities across interfaces.

5. Evaluation of geophysical wireline log data shows significant changes in dynamic elastic moduli throughout the Carmel Formation, a heterolithic caprock seal.
6. Numerical modeling allows exploration of mechanical stratigraphy and subsurface strain distribution. The variability in elastic properties across the reservoir/seal interface and across intra-seal interfaces influences local strain distributions and these changes in strain distribution are reflected in the density of fractures observed in outcrop.

7. Changing material properties affects failure mode/type, and analysis of the pore fluid factor shows that although each caprock seal analog reach similar maximum burial depths failure within the calcareous lithologies occurs as hydraulic extension, while the quartz-rich lithologies fail in the extensional shear regime.

8. At depth low cohesive strength rocks are very near failure and mechanical failure can be induced by small changes in pore fluid pressure.

**Implications**

Until recently, fine-grained clastic rocks have received little attention in the field of geomechanics. However, understanding the mechanical response of caprock seals to changing stress conditions is crucial for carbon capture and storage and in the effective production of unconventional resources. Analyses of appropriate geologic analogs are key in constraining the importance of heterogeneity to mechanical failure of cap-rock seals.
I summarize the implications of the enumerated conclusions above by combining Mohr-Circle failure criteria with burial history models for each locality to estimate the increase in pore fluid pressure required to cause mechanical failure of intact or cohesionless rock within the caprock as well as those that are critically stressed (Figure 6-1, Table 6-1).

This analysis shows that in all four analog studies the increase in pore-fluid pressure required for failure of a cohesionless material is at or very near hydrostatic pore-fluid pressure in all cases (Figure 6-1, Table 6-1). The fluid pressure requirement increases slightly with depth but is never more than 2 MPa over hydrostatic pore pressure (Table 6-1& 6-2). This has significant implications for management of subsurface storage sites specifically pressure and plume management in the impact area surrounding the active site of injection.

Using a numerical simulation provided by the Princeton Subsurface Hydrogeology Research Group CO2 Injection Simulator (http://monty.princeton.edu/CO2interface/) I model injection of 1 Mt/year of CO2 over 50 years, into a 150 m thick reservoir at 900m depth, 30% porosity, 0.05 Darcy permeability, and 30% residual brine saturation to estimate an idealized plume migration and associated pressure front. The 1 Mt/year injection rate is the injection flux for a single well that would be associated with a CO2 injection field of 10-20 wells, using CO2 captured from a 600 MW power plant, with approximately 6.6 Mt/year of CO2 emissions.
Figure 6-1. Fracture pressures required for mechanical failure above normal hydrostatic pore fluid pressure. Differential stress line plotted for reference. Data derived from Mohr-Coulomb failure analysis of rock properties derived from rock strength testing and burial history. Carmel Formation data from San Rafael Swell, and Organ Rock Formation from Hite Crossing.
Table 6-1. Carmel Formation failure pressures derived from burial history and Mohr-Coulomb failure analysis.

<table>
<thead>
<tr>
<th>Age (Ma)</th>
<th>Burial Depth (km)</th>
<th>ΔP_p for failure at C=0</th>
<th>Distance from Wellbore (km)</th>
<th>ΔP_p for failure at C=10.08</th>
</tr>
</thead>
<tbody>
<tr>
<td>44</td>
<td>4.43</td>
<td>1.37 MPa 198.70 psi</td>
<td>0</td>
<td>17.79 MPa 2580.22 psi</td>
</tr>
<tr>
<td>60</td>
<td>2.31</td>
<td>0.59 MPa 85.57 psi</td>
<td>&lt;5</td>
<td>16.99 MPa 2464.19 psi</td>
</tr>
<tr>
<td>84</td>
<td>1.63</td>
<td>0.38 MPa 55.11 psi</td>
<td>&lt;10</td>
<td>13.48 MPa 1955.11 psi</td>
</tr>
<tr>
<td>98</td>
<td>0.87</td>
<td>0.12 MPa 17.40 psi</td>
<td>&lt;20</td>
<td>9.49 MPa 1376.41 psi</td>
</tr>
<tr>
<td>147</td>
<td>0.77</td>
<td>0.11 MPa 15.95 psi</td>
<td>&lt;20</td>
<td>8.99 MPa 1303.89 psi</td>
</tr>
<tr>
<td>156</td>
<td>0.66</td>
<td>0.10 MPa 14.50 psi</td>
<td>&lt;20</td>
<td>8.45 MPa 1225.57 psi</td>
</tr>
</tbody>
</table>

Table 0-2. Organ Rock Formation failure pressures derived from burial history and Mohr-Coulomb failure analysis.

<table>
<thead>
<tr>
<th>Age (Ma)</th>
<th>Burial Depth (km)</th>
<th>ΔP_p for failure at C=0</th>
<th>Distance from Wellbore (km)</th>
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</thead>
<tbody>
<tr>
<td>44</td>
<td>3.53</td>
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<td>69</td>
<td>2.75</td>
<td>0.46 MPa 66.72 psi</td>
<td>&lt;10</td>
<td>8.46 MPa 1227.02 psi</td>
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<tr>
<td>84</td>
<td>1.95</td>
<td>0.38 MPa 55.11 psi</td>
<td>&lt;10</td>
<td>8.27 MPa 1199.46 psi</td>
</tr>
<tr>
<td>98</td>
<td>0.79</td>
<td>0.32 MPa 46.41 psi</td>
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<td>6.53 MPa 947.09 psi</td>
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<tr>
<td>146</td>
<td>0.72</td>
<td>0.14 MPa 20.31 psi</td>
<td>&lt;25</td>
<td>6.16 MPa 893.43 psi</td>
</tr>
</tbody>
</table>
The results of this model show that the pressure plume extends an order of magnitude farther from the injection point than that of the fluid plume. At 25 years post injection the pressure front exceeds the change in pore pressure necessary for failure of cohesionless material (Table 6-1, Figures 6-1 & 6-2). The pressure disturbance within the reservoir and adjacent to the reservoir seal interface could lead to mechanical failure within the overlying caprock seal and initiation of a high permeability pathway. At burial depths less than 2 km the change in pore-fluid pressure necessary to induce failure in both the siliciclastic Organ Rock Formation and the limestone rich Carmel Formation are below 0.4 MPa, which develops up to 10 km away from the injection point.

This simple numeric simulation combined with the failure pressure estimates that are dependent on rock type and lithostatic load show the importance of understanding the distribution of sub-seismic features to risk management in a sequestration scenario. A site-specific understanding of preexisting and past stress orientations, characterization of fault and fracture networks, and mechanical rock properties is necessary for risk assessment and management.

Based on the research presented in this dissertation the pressure increases associated with the fluid plume will likely encounter existing structural discontinuities and change the state of stress in the subsurface, at least locally. Encountering such cohesionless or critically stressed structures would result in failure of the CCS project due to loss of control of the fluid plume.
Figure 6-2. Plume evolution and pressure disturbance away from well bore at a time of 25 years after the start of injection. Top: plume evolution. Bottom: Pressure front evolution, vulnerable pressure zone highlighted for cohesionless material.
The accurate evaluation of pre-existing structures and of mechanical rock properties within caprock seals is important for successful storage of gas or fluid in the subsurface for appropriate storage safety and plume management. This dissertation shows the value of integrated field, laboratory, and simple modeling analyses to consider the nature of real rock properties and their responses to increased fluid pressures in engineered or naturally occurring systems.
APPENDICES
Appendix A: Chapter 1 Appendix
Figure A-1. Project timeline including: course work, teaching, field work and laboratory analysis, awards and grants, conference presentations and manuscript submissions/publications.
Appendix B: Chapter 2 Appendix
Table B-1. Borehole data used in this study

<table>
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<tr>
<th>Borehole Name</th>
<th>UWI</th>
<th>Wire line Logs Digitized</th>
<th>Logs Calculated</th>
<th>Depth Range MD/TVDSS meters (feet)</th>
<th>Core Available</th>
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<td>GR, DT</td>
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<td>Desert Lake U 5</td>
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<td>GR, DT</td>
<td>Vp, Vs, Den</td>
<td>236/1528 (774/5013)</td>
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<td>Desert Lake Unit 1</td>
<td>43-015-11328</td>
<td>GR, DT</td>
<td>Vp, Vs, Den</td>
<td>748/954 (2454/3130)</td>
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<td>Federal 41-33</td>
<td>43-015-30092</td>
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<td>Vp, Vs, Den</td>
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<td>Federal Mounds 1</td>
<td>43-015-10825</td>
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<td>Vp, Vs, Den</td>
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<td>Ferron Unit 5</td>
<td>43-015-20145</td>
<td>GR, DT</td>
<td>Vp, Vs, Den</td>
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<td>Vp, Vs, Den</td>
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<td>43-015-30338</td>
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<td>GR, DT</td>
<td>Vp, Vs, Den</td>
<td>1510/274 (4954/899)</td>
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</table>
B-2. Description of equipment used in this study

Schmidt Hammer N/L

The Schmidt hammer device measures a rebound value (R) from a spring-loaded pin mechanism that presses against the rock face. Specific relationships exist between the R-value and the hardness and compressive strength of the material. Impact direction must be taken into account when ascertaining R-values. Although commonly employed to measure the hardness and strength of concrete Selby (1993) presents a method for applying Schmidt Hammer use to geologic/geomorphic applications. This study employed the Selby method to measure R-values from natural rock. Test sites are more than 60 mm away from an edge or joint, and test surfaces are clean smooth and planar. The surface is impacted 5 times and the highest reading is used. The hammer is held perpendicular to the rock face. In this study R-values are used as an index for strength and not converted to uniaxial compressive strength (MPa).
TinyPerm II

TinyPerm II is a portable hand-held air permeameter; it can be used for measurement of rock matrix permeability or effective fracture aperture on outcrop and core (NER, 2012). Measurements were taken on a clean surface 60 mm away for an edge or discontinuity, and the lowest permeability value is reported. For intact rock, TinyPerm II ranges from 10 mDarcy to 10 Darcy. The TinyPerm II field measurements were cross-referenced to the calibration curve to obtain absolute permeability. The equation for the calibration curve is: \( T = -0.8206 \times \log_{10}(K) + 12.8737 \) , where \( T \) is the TinyPerm II measurement and \( K \) is the permeability in Darcy (D). The conversion factor for D to \( m^2 \) is \( 1 \text{ D} = 9.869233 \times 10^{-13} \text{ m}^2 \)

B-3. Equations used in this study

Equations 1 and 2 presented below are those used for conversion of sonic log travel time to velocity (Equation 1), and sonic log travel time from density log data or vise versa (Equation 2). Equations 3 and 4 are those used to calculate elastic moduli from the wireline well log velocity data. Equation 3 gives Poisson’s ratio (dynamic) and Equation 4 gives Young’s modulus (dynamic).

Equation 1 – sonic log travel time to velocity \((V)\)

Travel time (ms/ft)

\[ 1 \times 10^6 \text{ s/1 ms} = V \text{ (ft/s)} \]

Converting to \( V \text{ (km/s)} \)
1 ft/1 s = 3.048*10^{-4} \text{ km/s} \\

Equation 2 – estimated sonic log travel time (ms/ft) from density log (g/cc) \\

Travel time (ms/ft) = \frac{1 \times 10^6}{357.3458 \times \text{density value}^4} \\

Equation 3 - Poisson’s ratio (dynamic) \( n_d \) (McCann and Entwisle, 1992) \\

\[
\nu_d = \frac{1}{2} \left[ \frac{V_p}{V_s} \right]^2 - 2 \\
\]

Equation 4 - Young’s modulus (dynamic) \( E_d \) (McCann and Entwisle, 1992) \\

\[
E_d = 2r(Vs^2)(1-n_d)
\]
Appendix C: Chapter 3 Appendix
Table C-1. Weight percent iron from elemental analysis obtained from ICPM90A. Samples compare iron content of bleached and unbleached fracture zone samples within the Organ Rock Formation.

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<th>Organ Rock Formation</th>
<th>wt. % Fe</th>
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<td>gc-or-1ub</td>
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<tr>
<td>unbleached host rock</td>
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<tr>
<td>gc-or-1b</td>
<td>1.08</td>
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<tr>
<td>bleached fracture margin</td>
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<tr>
<td>gc-or-7ub</td>
<td>1.33</td>
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<tr>
<td>unbleached host rock</td>
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<tr>
<td>gc-or-7b</td>
<td>0.93</td>
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<tr>
<td>bleached fracture margin</td>
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Table C-2. Elemental analysis data obtained from ICPM90A. Data table organized by major, trace, and rare earth elements then atomic number. Red columns contain data from hematite veins; blue columns contain data from calcite veins.

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<th>ELEMENT</th>
<th>DETECTION</th>
<th>PROTOLITH</th>
<th>FAULT ZONE HOST ROCK</th>
<th>FAULT ZONE VEIN FILL</th>
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<td>(units)</td>
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Table C-3. Mineralogy from x-ray powder diffraction. (+) Common; (--) Trace; (-) Rare

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<th>MICROCLINE</th>
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<th>ORTHOCLOISE</th>
<th>HEMATITE</th>
<th>GEOTHITE</th>
<th>WÜLFELITE</th>
<th>WUSTITE</th>
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<th>BARATONITE</th>
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Appendix D: Chapter 4 Appendix
Figure D-1. Organ Rock Formation burial history curve (top), Mohr circles and Coulomb-Griffith envelopes for failure of C=0 and C=6.28 MPa (bottom).
Figure D-2. Carmel Formation burial history curve (top), Mohr circles and Coulomb-Griffith envelopes for failure of C=0 and cohesive strength C=10.06 MPa (bottom).
Figure D-3. Earthy Member Entrada Formation burial history curve (top), Mohr circles and Coulomb-Griffith envelopes for failure of $C=0$ and cohesive strength $C=4.74$ MPa (bottom).
Figure D-4. Tununk Member burial history curve (top), Mohr circles and Culomb-Griffith envelopes for failure of $C=0$ and $C=21.02$ MPa.
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**Note:** $\lambda_v = \frac{\Delta P_p}{P_p - \sigma_1}$ for intact rock.
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|-----------|---------|------------|-----------------|-----------------|-----------------|------------------------|--------------------------|--------------------------|--------------------------|--------------------------|--------------------------|--------------------------|----------------|-----------|
| Earthy Member | 146.0 | 0.72 | 18.39 | 10.81 | 7.58 | 7.07 | 5.16 | 11.18 | 13.23 | 7.21 | 6.16 | 0.14 | H.E. | 0.72 | 0.39 |
| | 98.0 | 0.79 | 20.28 | 11.92 | 8.36 | 7.80 | 5.95 | 12.16 | 14.33 | 8.12 | 6.53 | 0.32 | H.E. | 0.71 | 0.40 |
| | 84.0 | 1.95 | 49.70 | 29.21 | 20.49 | 19.12 | 22.17 | 30.23 | 27.53 | 19.47 | 8.41 | 0.35 | E.S | 0.55 | 0.39 |
| | 68.5 | 2.75 | 70.34 | 41.34 | 29.00 | 27.05 | 34.82 | 42.82 | 35.52 | 27.52 | 8.46 | 0.46 | E.S | 0.50 | 0.39 |
| | 53.0 | 3.42 | 87.45 | 51.39 | 36.06 | 33.63 | 44.69 | 52.66 | 42.76 | 34.79 | 9.12 | 1.15 | E.S | 0.49 | 0.40 |
| | 44.0 | 3.53 | 90.23 | 53.03 | 37.20 | 34.70 | 45.03 | 53.99 | 45.20 | 36.24 | 10.50 | 1.54 | E.S | 0.50 | 0.40 |
| Tununk Member | 84.0 | 1.15 | 29.42 | 17.29 | 12.13 | 11.32 | 1.57 | 17.8 | 27.85 | 11.62 | 16.54 | 0.31 | H.E. | 0.95 | 0.40 |
| | 68.6 | 1.96 | 50.06 | 29.42 | 20.64 | 19.25 | 10.14 | 30.32 | 39.92 | 19.74 | 20.66 | 0.48 | H.E. | 0.80 | 0.64 |
| | 53.0 | 2.63 | 67.17 | 39.48 | 27.69 | 25.83 | 17.14 | 40.7 | 50.03 | 26.47 | 24.20 | 0.64 | H.E. | 0.74 | 0.64 |
| | 44.0 | 2.74 | 69.95 | 41.11 | 28.84 | 26.91 | 18.35 | 42.24 | 51.60 | 27.71 | 24.70 | 0.81 | H.E. | 0.74 | 0.64 |
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I am in the process of preparing my written dissertation as part of completion of my Doctoral degree at Utah State University. I am writing seeking your permission to include the Journal of Structural Geology paper you were listed as a co-author on for one of my dissertation chapters.

Please indicate your approval of this request by signing the letter where indicated below and returning a copy of this letter to me via email, fax (435) 797-1588, or mailed directly to the geology department address listed below.

Thank you

Elizabeth Petrie
PhD Candidate
USU Department of Geology
4505 Old Main Hill
Logan, UT 84322-4505
espetrie@aggiemail.usu.edu

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Permission Granted for the Use Requested Above:

By:
Name: Robert A. Petrie
Signature: [Signature]

Title: Manager Eastern Hemisphere Baroid IOP
Date: 12/30/13
December 30, 2013

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I am in the process of preparing my written dissertation as part of completion of my Doctoral degree at Utah State University. I am writing seeking your permission to include the submitted manuscript to the AAPG Bulletin on which you are as a co-author as one of my dissertation chapters.

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Thank you

Elizabeth Petrie
PhD Candidate
USU Department of Geology
4505 Old Main Hill
Logan, UT 84322-4505
espetrie@aggiemail.usu.edu

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By:
Name: Stephen J. Bauer

Signature:  
Title: Distinguished Member of the Technical Staff, Geomechanics, Sandia National Laboratories

Date: 12/30/13
December 30, 2013

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I am in the process of preparing my written dissertation as part of completion of my Doctoral degree at Utah State University. I am writing seeking your permission to include the Environmental Geosciences and Journal of Structural Geology publications as well as the manuscripts for the American Association of Geologists Bulletin and Bulletin of Canadian Petroleum Geology you were listed as a co-author on as dissertation chapters.

Please indicate your approval of this request by signing the letter where indicated below and returning a copy of this letter to me via email, fax (435) 797-1588, or mailed directly to the geology department address listed below.

Thank you

Elizabeth Petrie  
PhD Candidate  
USU Department of Geology  
4505 Old Main Hill  
Logan, UT 84322-4505  
espetrie@aggiemail.usu.edu

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Permission Granted for the Use Requested Above:

By:  
Name: James P. Evans  
Signature: [Signature]

Title: [Title]
Cirriculum Vita
Education

**Ph.D., Geology**, Utah State University (USU), Logan, UT, Defended 11/2013, degree date 05/2014
*Rock strength of cap rock lithologies: evidence for past seal failure, migration of fluids and analysis of the reservoir-seal interface in outcrop and the subsurface.*

**M.S., Geology**, USU, Logan, UT, 2003
*Sequence stratigraphic correlation of the Wheeler Formation: a geochemical approach.*

**B.S., Earth and Planetary Sciences**, *cum laude*, University of New Mexico (UNM), Albuquerque, NM, 2000
*Undergraduate Research Thesis: Depositional environments and diagenesis of Neo-Proterozoic Chuar Group carbonates.*

**B.S., Biology**, UNM, Albuquerque, NM, 2000

Research and Teaching Experience

**Research Assistant – USU**
2010-present
Rock mechanics and seal integrity. Includes: outcrop characterization, fracture analysis, optical petrography, X-Ray diffraction, inductively coupled plasma mass spectrometry, tri-axial tests, unconfined compressive strength, wireline log analysis, numerical and geomechanical modeling

**Instructor – USU**
2012-present
*Geology 3200* – Historical geology lecture
*Geology 3700* – Structural geology lecture (emergency substitute instructor)
*Geology 5200* – Field camp structural geology project
*Geology 3200* – Historical geology online flex course development and instruction

**Teaching Assistant - USU**
2011
*Geology 5530* Petroleum Systems: Principals of Exploration and Development
Mentor of USU’s Imperial Barrel Award Team, Rock Mountain Region IBA Champions
*Geology 5560* Subsurface Mapping: Principals and Techniques.
Teaching Assistant - USU  
2000-2002  
Geology 1150 Introductory geology for science majors  
Geology 3200 Earth History  
Geology 3550 Sedimentology and Stratigraphy  
Geology 5200 Field Camp  

Research Assistant - UNM  
1998  

Employment History  

Chief Geologist, Drillsearch Energy Ltd., Sydney, NSW, AUSTRALIA  
2005-2009  
Responsible for management of exploration, production, and operational activities including:  
**Geology and geophysics**  
Manage and review, or carry out geologic and geophysical analysis in operated and non-operated areas; seismic and wireline log interpretation, data management, management of contract geologist and geophysicist, create and maintenance of project budgets and cash flows.  
**Joint Venture Representative**  
Attend and/or hold management and technical committee meetings; evaluate/create proposed work programs; liaison for JV permit matters and Australian Stock Exchange announcements  
**Business Development**  
Farm in/out – Preparation and coordination of farm out opportunities; evaluate/advise on farm in offers received  
Acreage gazettal – evaluate and prepare government bid applications  
**Government Correspondence** –quarterly and annual permit reports; ensure technical compliance with regulations  
**Corporate Responsibilities**  
Prepare investor and industry conference presentations and media material. Draft technical data for market releases including daily drilling reports, operations update, quarterly, half-yearly and annual reports.  

Senior Geologist, ExxonMobil Exploration Corp., Houston, TX  
2003-2005  
Sequence stratigraphic correlations between mature heavy oil fields in Western Canada.  
3D seismic interpretation and well log analysis collaborative study between ExxonMobil and Pemex. Fluency in Spanish assisted in communication between the business units involved.
3D volume interpretation and visualization. Company-wide service work for time-limited 3D seismic volume interpretation projects, well planning, and geologic modeling projects using Voxelgeo and Gocad based platforms. Taught other geoscientists 3D volume interpretation techniques, shortening interpretation turn around time via one-on-one mentoring, 2-5 day classes, and hosting visualization demos throughout the exploration and development companies.

**Geologist, U.S. Geological Survey, Menlo Park, CA**
**Summers 2000-2001**
Two field seasons in boreal forests of Canada and Alaska. Data collection for quantification of the effects of forest fire disturbance and subsequent successional processes on the exchange of energy, water, and nutrients between the land surface and atmosphere. Soil and down woody debris sampling in Canada (NWT and Manitoba) and Alaska. Responsible for reconnaissance fieldwork and preparation of field sites for 5-year study, preliminary soil sampling and descriptions, down woody debris sampling, soil organic carbon analysis and stable isotope analysis.

**Geochemistry Lab Technician, UNM, Earth & Planetary Science, Albuquerque, NM**
**1998-2000**
Conducted sample preparation and analysis for students and faculty within the department of E&PS as well as Sandia National Laboratories, Los Alamos National Laboratories, the Better Business Bureau, and private consumers, using AAS, XRD, ICP-MS, and iron-titration.

**Geologist, Getchell Gold Mine, Golconda, NV**
**Summer 1998**
Core-shed geologist at mining operation in northwestern Nevada. Described, split, and prepared core and cuttings for assay and archiving.

**Funded Proposals**

**2013** Shell Global Solutions Research Grant: Cedar Mesa Sandstone reservoir characterization and association of structural discontinuities and diagenesis [$50,000].

**2011** Grant for Geology Department purchase of Cavity Ring Down Spectrometer for measurement of stable carbon isotope ratios [$93,000]

**2011** GDL Foundation Structural Diagensis Fellowship: Evaluation of discontinuities and diagenesis in cap-rock lithologies; and implications for secure storage of CO$_2$ [$8,000]

**2002** Geological Society of America, Graduate Student Research Grant [$500]

**2002** AAPG Grants-in-Aid, Kenneth H. Crandall Memorial Grant [$500]

**2002** Colorado Scientific Society Research Grant, Steven Oriel Memorial Fund [$420]

**Scholarships and Awards**

**2013** USU Geology Department Robert Q. Oaks Citizenship Award [$200]

**2011** USU Geology Department Outstanding Graduate Researcher [$500]

**2002** USU Graduate Studies Merit Scholarship Spring Semester [$500]
2002 Utah State University Outstanding Graduate Teaching Assistant [$500]
2002 Utah State University, Department of Geology, J. Stewart Williams Fund [$500]
2000 V. C. Kelley/Estwing Outstanding Field Geologist Award
1999 Harry and Mabel Leonard Scholarship [$1,500]
1999 Lucille Pipkin Undergraduate Scholarship [$500]

Publications

PETRIE, E.S., J.P. Evans, Modeling the significance of mechanical interfaces to fracture propagation and morphology – Geomechanical models derived from outcrop analysis. IN PREP: Bulletin of Canadian Petroleum Geology

PETRIE, E.S., J.P. Evans, S.J. and Bauer, Failure of caprock seals as determined from mechanical stratigraphy, stress history and tensile failure analysis of exhumed analogs. ACCEPTED American Association of Petroleum Geologists Bulletin


PETRIE, E.S., T.N. Jeppson, J.P. Evans, 2012, Predicting rock strength variability at stratigraphic interfaces in caprock lithologies at depth: correlation between outcrop and subsurface: Environmental Geosciences,19 (4), 125-142.

Presentations


PETRIE, E.S., J.P. Evans, 2013, Modeling the significance of mechanical interfaces to fracture propagation and morphology – Geomechanical models derived from outcrop analysis. Unconventional Resources Technology Conference, URTeC Control ID Number: 1619873.


PETRIE, E. S., J.P. Evans, T.N. Jeppson, 2011, Fracture behavior across interfaces in seal lithologies, *invited*. Abstract V13E-01 oral presentation AGU Fall Meeting


Professional Organizations

Geological Society of America
American Association of Petroleum Geologists
Society for Exploration Geophysicists
American Geophysical Union
Rocky Mountain Association of Geologist
AAPG Prowess Committee