Fracture distribution on the Swift Reservoir Anticline, Montana: implications for structural and lithological controls on fracture intensity

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Abstract

Where primary porosity and permeability of a rock are unfavourable for hydrocarbon production, fractures can improve reservoir potential by enhancing permeability. Higher fracture intensity may create a better connected fracture network, improving fractured reservoir quality. Investigations into the controls on fracture intensity commonly conclude that either structural or lithological factors have the greatest influence on fracture abundance. We use the Swift Reservoir Anticline in north western Montana to investigate how fracture intensity varies throughout the structure, and determine that although structural factors do influence fracture intensity, lithology is the main control at outcrop.

The Swift Reservoir Anticline exposes bedding surfaces of the Mississippian Castle Reef Formation dolomite. Field data indicates that fracture intensity is highest in the fold forelimb, decreasing into the backlimb except in outcrops of coarse dolomite where fracture intensity is low, regardless of structural position. Field fracture intensity correlates with whole rock quartz, kaolinite and porosity percentages. We suggest porosity and composition influence bulk rock mechanical properties, which, in turn, control the fracture intensity at outcrop. Fracture intensity has a stronger relationship with lithological than structural factors, therefore we suggest that the key to predicting fracture intensity in the subsurface here is understanding how lithology varies spatially.

Introduction

Fractured reservoirs host hydrocarbon reserves globally, and are particularly beneficial to petroleum systems with little primary porosity and permeability. Fractured reservoirs are productive from a wide range of lithologies, from sedimentary rocks such as the tight sandstones of Bolivia (Florez-Niño et al., 2005; Iñigo et al., 2012; Heidmann et al., 2017), to basement rocks such as the fractured granites of Vietnam (Cuong & Warren, 2009). These reservoirs are found in a range of geological settings, from relatively undeformed regions, where fractures form in response to regional stresses or increased pore pressure (e.g. Engelder & Lacazette, 1990; Lacazette & Engelder, 1992), to tectonically deformed regions, where stress concentrations have formed tectonic fractures on folds and around faults. Fractured reservoirs form a significant contribution to global oil and gas reserves. For example the Zagros fold-thrust belt, where oil and gas are produced primarily from the fractured Asmari Formation carbonates, is one of the most prolific onshore regions of hydrocarbon exploration and production. It
has been estimated that this region alone contains 49% of the global fold-thrust belt reserves (Cooper, 2007).

The contribution of fracture networks to the petroleum system varies depending on several fracture attributes. Wider fracture apertures increase the secondary porosity and permeability, increasing fluid flow (Odling et al., 1999); longer fractures increase the likelihood of fracture intersection, and therefore can improve fracture network connectivity. Fracture network orientations play a role in influencing fluid flow; orthogonal fracture sets are likely to intersect, improving fracture connectivity and increasing the size of the effective fracture network (see Watkins et al., 2015a). The intensity of fracture networks can also control fluid flow; higher fracture intensities (the total fracture length, area or volume within a given area or volume) correlate with higher fracture connectivity (e.g. Watkins et al., 2018); if a fracture network is well connected, fluids are able to migrate through the fractures more easily. The controls on the spatial distribution of fracture network intensity is not fully understood. A better understanding of the controls on fracture intensity distribution would allow better prediction of the highest quality fractured reservoirs in the subsurface, and therefore better targeting of these permeability sweet spots when drilling for oil and gas.

Current understanding of the controls on fracture intensity variation is often derived from outcrops used as analogues to subsurface fractured reservoirs. These studies suggest that both structural and lithological controls influence fracture intensity (e.g. Fischer & Jackson, 1999). Examples from a range of sedimentary lithologies include studies that attribute fracture intensity variation to bed thickness changes (e.g. Hobbs, 1967; Gross et al., 1995; Wennberg et al., 2007; Ortega et al., 2010; Barbier et al., 2012), composition (e.g. McQuillan, 1973; Hugman & Friedman, 1979; Corbett et al., 1987; Hanks et al., 1997; Ferrill & Morris, 2008), grain size (e.g. Hanks et al., 1997; Nelson, 2001; Wennberg et al., 2007), porosity (e.g. Corbett et al., 1987; Barbier et al., 2012), ‘degree of tectonic deformation’ (e.g. Hobbs, 1967), structural position on a fold (e.g. Hanks et al., 1997; Wennberg et al., 2007), proximity to faults (e.g. Hanks et al., 1997), and bedding curvature (e.g. Ramsay, 1967; Lisle, 1992; Lisle, 1994; Ortega et al., 2010). The question is, in a geological setting where any number of these variables change, which are the dominant factors that control fracturing?

Evidence for structure and lithology controlling fracture intensity are well reported in the literature (see previous paragraph); we aim to add to this knowledge base, assessing the relative contribution of lithological and structural controls on fracture formation using an anticline of dolomite in the Sawtooth Range of Montana, USA. Using field data we separately assess how structural (simple curvature and bedding dip) and lithological (grain size, porosity, composition) factors correlate to fracture intensity variation, and we use results from this analysis to discuss which has the greatest influence on fracture formation.

Sawtooth Range

Regional Geology

We use an anticline in the Sawtooth Range of Montana, USA, to analyse fractures. The Sawtooth Range is an arcuate fold-thrust belt formed in the Palaeocene (Mudge, 1982) during the Cordilleran Orogeny (Fuentes et al., 2012). It is located on the eastern edge of the Rocky Mountains in north-west Montana, bounded to the east by the Mesozoic-Palaeogene foreland basin, and to the west by the Lewis and Eldorado Thrust system, which separates Proterozoic-Palaeozoic rocks in its hangingwall from Palaeozoic-Mesozoic rocks in its footwall (Figure 1). The Sawtooth Range is in the footwall to the
Lewis-Eldorado Thrust system, and has been interpreted by many authors to be a thin-skinned fold-thrust belt (Mudge, 1982; Mitra, 1986; Holl & Anastasio, 1992; Fuentes et al., 2012).

The Sawtooth Range has been chosen for this study to represent an along-strike equivalent to the subsurface fractured reservoirs in the Front Ranges of Alberta. Here, most hydrocarbons are produced from Carboniferous carbonates similar to those exposed in the Sawtooth Range; some intervals rely on fracture permeability for commercial production (e.g. Rawnsley et al., 2007). The Alberta Foothills reservoir equivalent intervals and trap geometries, however, are not well exposed at outcrop. In contrast, equivalent carbonates and anticlinal trap geometries can be observed on well-exposed outcrop in the Sawtooth Range, and as such they make good outcrop analogues to study the causes behind variation in fracture networks. For this study we focus on one anticline, namely the Swift Reservoir Anticline. The present erosion level exposes extensive areas of bedding surfaces associated with several depositional cycles near and at the top of the Mississippian. All observed bedding surfaces on the Swift Reservoir Anticline are heavily fractured, so the spatial distribution of fracture intensity variation in 2D can be studied. However, for the outcrops studied only the top surface of beds can be observed, meaning fracture network variation in 3D cannot be studied, nor can the control of mechanical layer bedding thickness on fracture intensity. In this study we aim to determine how structural controls relating to the geometry of the fold, as well as lithology, influence fracture intensity by analysing fracture data from various structural positions, and from the tops of different beds with different lithological properties.

**Structural, Burial and Regional Stress History**

NW Montana and adjacent SW Alberta record a complex tectonic and depositional history from PreCambrian to Cenozoic time (e.g. Mudge, 1982; Price, 1994; Fuentes et al., 2012). We only address here components potentially relevant to burial history and brittle deformation of the Lower Mississippian rocks observed at Swift Reservoir Anticline. Lower Mississippian rocks were deposited on a continental margin most commonly interpreted as an eastward-tapering marine foreland or back-arc basin that developed in response to convergent plate motion to the west. Early Mississippian deposition was influenced by local margin-parallel extensional faults perhaps related to foreland basin flexural processes or back arc extension, and by reactivation of pre-existing structures oblique to the margin (Reid and Dorobek, 1993; Batt et al., 2008; Cooley et al., 2011). A significant Carboniferous to Jurassic unconformity in the Sawtooth Range records widespread non-deposition and/or erosion in NW Montana and SW Alberta, during ongoing convergent plate interactions to the west. Much of this unconformity is perhaps associated with passage of a Jurassic flexural forebulge prior to local thrusting, driven by a period of terrane accretion and thrusting to the west (Ward and Sears, 2007; Fuentes et al., 2011). This unconformity is well exposed on Swift Reservoir Anticline (Mudge, 1982; Ward & Sears, 2007). Therefore the pre-Cretaceous tectonic history implies possibly complex perturbations in stress regime with related brittle deformation during and following deposition, but before compressional folding. Pre-Jurassic deformation at Swift Reservoir is addressed in more detail below.

By Middle Jurassic time, a north-eastward propagating fold-thrust belt developed on the former continental margin, accompanied by widespread subsidence and clastic fill of an associated foreland basin to the east. Fold-thrust deformation, including the Swift Reservoir Anticline, involved the Paleozoic carbonate and Mesozoic clastic rocks of the Sawtooth Range by latest Cretaceous to Paleocene time (Fuentes et al., 2012).

Extension occurred in NW Montana during the mid-Eocene to Oligocene (Constenius, 1996; Fuentes et al., 2011) in association with regional, crustal-scale extension in the hinterland parts of the thrust
There is little or no stratigraphic record or clear structural evidence for extensional deformation in the Montana Sawtooth Range or correlative southern Alberta Foothills; however, Eocene to Oligocene syn-extensional clastic deposits are locally exposed in extensional fault hangingwalls to the northwest of the Sawtooth Range (McMannis, 1965; Constenius, 1996; Figure 1). Therefore we cannot rule out the possibility that the extensional event contributed to fractures observed at Swift Reservoir Anticline; there is however no direct evidence for this, and present-day maximum horizontal stress orientations are consistently SW-NE as supported by abundant subsurface data in central to southernmost Alberta (Bell and Babcock, 1986; Bell, 1994; Heidbach et al., 2008). Burial data for subsurface Mississippian rocks in the eastern Foothills in southern Alberta indicate increasing sedimentary and subsequent tectonic burial from the Jurassic to a maximum of close to 10 km in the Paleocene. This was followed by exhumation to their current depth of on the order of 5 km (Price, 1994; Hardebol et al., 2009). In the Sawtooth Range equivalent rocks were probably buried much less deeply before Cenozoic uplift and present exposure at surface. Quantitative data however are limited to perhaps uncertain conodont colour alteration index analyses from the Mississippian carbonate section that imply a maximum burial to no more than about 2-3 km (Nichols, 1986).

**Mississippian Stratigraphy**

Cyclicity and depositional thickness variations in the Lower Mississippian Castle Reef Formation contribute directly to the mechanical heterogeneity of the carbonate succession and thus fracture distribution at Swift Reservoir Anticline. The Lower to Middle Mississippian Madison Group contains two main lithostratigraphic subdivisions in the Sawtooth Range (Mudge, 1962; Nichols, 1984): the lower, dominantly limestone Allan Mountain Formation and the conformably overlying Castle Reef Formation, comprising mainly bioclastic dolomites with both primary and secondary dolomitization. The Castle Reef Formation has in turn been divided into a lower Gateway Pass Unit of crinoidal grainstones and interbedded dolomitic mudstone/wackestone, grading upwards into a Dupuyer Creek Unit of cyclically interbedded packstone, wackestone and mudstone, which is well exposed at Swift Reservoir Anticline (Nichols, 1984). The Castle Reef Formation varies in thickness due to erosion beneath the sub-Jurassic unconformity over the entire Sawtooth Range but averages 215 m in the Swift Reservoir area (Singdahlsen, 1986), and the Dupuyer Creek Unit varies in thickness from 30-75 m.

The entire Mississippian succession in western Montana is interpreted (Reid and Dorobek, 1993; Batt et al. 2007) to record overall a prograding carbonate ramp from deep water (Allan Mountain Formation) through shallow high-energy conditions (Castle Reef Formation, Gateway Pass Unit) to a tidally-influenced interior ramp setting (Castle Reef Formation, Dupuyer Creek Unit). Deposition in the upper, low-accommodation part of the section is marked by higher-order sea-level fluctuations, resulting in lithological cyclicity, on the order of 0.5 to 5 m, from mud-dominated to grain-dominated facies in the Dupuyer Creek Unit where it is exposed at Swift Reservoir (Figure 2). The current erosion surface across the Swift Reservoir Anticline exposes approximately 100 m of Castle Reef Fm. beneath the unconformity, so it is likely that the coarser-grained facies at the lowest exposure levels are within the upper part of the Gateway Pass Unit.

**Swift Reservoir Anticline**

The Swift Reservoir Anticline is on the north-eastern margin of the Sawtooth Range (Figure 1, Figure 3a), bounded to the west by structurally overlying fold-thrust structures and to the east by less deformed Mesozoic rocks in the foothills and western foreland basin (Figure 3d). The Swift Reservoir Anticline exposes folded Mississippian Castle Reef Formation dolomites, which form a strongly
asymmetric fold structure (Figure 3b, 3d). The fold has a steep forelimb dipping to the northeast, and shallow backlimb dipping to the southwest. The thrust relationships in this part of the Sawtooth Range are very complex and are best illustrated by the detailed surface maps of Ross (2016), north of Swift Reservoir, and Singdahlsen (1986), south of Swift Reservoir. These maps have been used together with data collected during this project to create the cross-section through the Swift Reservoir Anticline (Figure 3d). An additional critical data constraint for the structural model is the Blackfeet Tribal 12-1 well. The well data publicly available from the Montana Bureau of Mines and Geology is limited to the location, total depth and the depths of the three penetrations of the top of the Mississippian, but Singdahlsen (1986; cross-section A-A') indicates that he had access to additional significant stratigraphic detail in the well.

The Blackfeet Tribal 12-1 well is located on the outcropping Cambrian sediments in the hangingwall of the Major Steel Backbone Thrust (MSBT). The well penetrates the underlying thrust sheet that includes the Swift Reservoir Anticline, and it intersects the Swift Thrust close to the base of the Mississippian section. In the footwall of the Swift Thrust, a structurally thickened package of the Cretaceous Kootenai Fm correlates up-dip to surface to the Eagle Creek Imbricate Zone (ECIZ) (Singdahlsen, 1986) with a floor thrust, the Fish Lake Thrust (FLT), located close to the base of the younger Cretaceous Blackleaf Fm. Beneath this floor thrust, a conformable sequence from the Blackleaf Fm to the base of the Mississippian is carried by the deepest thrust in the system, the Old Man Thrust (OMT). The Mississippian in the hangingwall of the OMT outcrops to the south (Berg, 2002 and Singdahlsen, 1986), confirming the subsurface geometry implied by the well. The well penetrates the OMT with Cretaceous Kootenai Fm in the footwall, and ultimately terminates in the upper Mississippian which is apparently at regional elevation and likely unthrusted.

Although the model is well constrained by surface and subsurface data, uncertainty remains for the evolution of the Swift Reservoir Anticline, mostly due to out-of-sequence thrusting. The OMT thrust sheet when restored would contain at its western end the footwall ramp cut-off that matches the hangingwall cut-off represented by the forelimb of the Swift Anticline. This is not shown on the cross section (Figure 3d) because the MSBT appears to be out-of-sequence, truncating two thrust repeats of Mississippian stratigraphy in its footwall. The OMT folds the FLT and likely folds the Swift Thrust as well. The OMT, close to the hangingwall cut-off of the Mississippian, is imbricated by a couple of thrust splays one of which comes to surface very close to the FLT. Another splay is mapped as the Mitten Lake Thrust (MLT). The final splay is the most external thrust mapped which is not named but is labelled as A on the cross section (Figure 3d). It is therefore difficult to determine the kinematic history of the Swift Reservoir Anticline given the structural complexity described above, particularly the lack of a matching footwall counterpart to the anticline. A suggested model of early development, however, is shown in Figure 3e. This model proposes an early history as a fault propagation fold, perhaps combined with trishear type strain ahead of the propagating fault tip, which would result in the tight fold hinge and steep to vertical or locally overturned forelimb.

To study how the fracture network varies on the Swift Reservoir Anticline, first we need to be sure that the fractures observed at outcrop might relate to the Late Cretaceous-Palaeocene compressional folding process rather than another earlier or later phase of deformation. Karst-widened fractures associated with the Carboniferous-Jurassic uplift can be found in the Castle Reef Formation dolomite at Swift Reservoir (Ward & Sears, 2007). These fractures penetrate the Castle Reef Formation for up to 4 m beneath the unconformity surface, and are filled with cherty sandstone and conglomerate from the base of the Jurassic (Ward, 2007; Ward & Sears, 2007). However we only observed evidence for these fractures in the fold forelimb at the northern end of the structure close to the preserved unconformity surface; elsewhere no evidence for karstified fractures could be seen. The outcrops
containing the karstified fractures also contain narrow aperture, non-karstified joint sets that are alike in attributes to the rest of the fractures that we sampled throughout the fold. These narrow fractures cut across the sand and conglomerate fill in the karstified fractures so are clearly younger. It is these narrow, younger fractures that we have sampled in this study; although older fracture sets are present in the structure, they are easily distinguished in the field due to their sand/conglomerate fill, and have been excluded from this study.

Methods

Fieldwork

To assess both the structural and lithological controls of fracture intensity on the Swift Reservoir Anticline we collected data in the field from eleven transects oriented normal to the fold hinge along a 2 km segment of the fold (Figure 4). The lengths of individual transects ranged from 145-560 m and the spacing between transects ranged from 95-430 m, depending on outcrop availability. Field data was collected from a total of 193 sampling sites, which were located along transects in fold backlimb, hinge and forelimb positions. At each sampling site the bedding dip was measured, a scaled and oriented photograph was taken orthogonal to bedding, and the lithology was recorded. All sampling sites are located in the Castle Reef Formation dolomite; lithology was classified depending on the grain content of the rock. Sampling sites were allocated one of three lithological classifications: 1) Grain Supported Dolomite (GSD); coarse dolomite (grainstone) with shell and coral fragments, 2) Mud Supported Dolomite (MSD); fine grained dolomite (mudstone) lacking clear grains and an intermediate phase of 3) Mud/Grain Supported Dolomite (MGSD); fine grained dolomite (wackestone-packestone) with identifiable grains and small shell fragments. Variations in lithology are attributed to sampling sites being located on different bedding surfaces (i.e. from different stratigraphic horizons in the Castle Reef Formation). As well as classifying lithology in the field, hand specimens from 10 of the sampling sites (5 GSD & 5 MSD) were collected for further analysis (see Figure 4 for hand specimen locations).

Fracture intensity estimation

Fracture (joint) intensity was estimated for each sampling site using digital circular scanlines. Bedding surface photographs taken at each sampling site were scaled and oriented using Move software. A digital circle of known radius was placed on the photograph and the number of intersections with this circle was recorded. The fracture intensity for each sampling site was estimated using Mauldon’s circular scanline method:

\[ I = \frac{n}{4r} \]

\( I \) = estimated fracture intensity (m/m²), \( n \) = number of fracture intersections with the digital circle, and \( r \) = circle radius (m) (Mauldon et al., 2001). Estimated fracture intensity is given as fracture length per unit area on the bedding surface. Sampling circle radii ranged from 8-39 cm; the exact size of the circle was chosen to attain a minimum \( n \) value of 30, as suggested by Rohrbaugh et al. (2002). Circular scanline sampling was chosen because it does not incur any orientation bias, and therefore no orientation correction is required. Although the circular scanline method only estimates fracture intensity it has been tested against other data collection methods that record the actual fracture intensity at outcrop, such as areal sampling (Watkins et al., 2015b), and has been shown to produce accurate results.

Lithological analysis
From the 10 hand specimens oriented thin sections for each were cut normal to bedding and parallel to the bedding dip direction. SEM (Scanning Electron Microscope) analysis was used to analyse thin sections using Back Scatter Electron (BSE) imaging. 12-15 randomly selected images of each thin section were collected for further image analysis using ImageJ software. In ImageJ, the percentages of lithological properties (porosity, dolomite, quartz, calcite, kaolinite, other mineral) were calculated for each image using the following workflow: 1) set image scale, 2) adjust image brightness and contrast until the lithological property of interest is shown in black, and the rest of the image is shown in white, 3) ‘despeckle’ the image to remove noise, 4) convert image to binary, 5) analyse the binary image to determine the percentage area of the lithological property of interest.

**3D model building & predictions**

Pseudosurfaces for the top Castle Reef Formation and fifteen arbitrary horizons beneath were constructed using Move software to extract curvature data from. These surfaces were built to represent the large scale fold geometry rather than the actual top Castle Reef Formation, which may, in reality, not be a single horizon surface. Cross sections for the Top Castle Reef Formation were constructed for each transect using field bedding data and geological map boundaries (Mudge & Earhart, 1983). A 3D surface for this horizon was then constructed using a spline curves algorithm; the resultant surface was analysed using curvature analysis to determine non-geological anomalies on the surface geometry. The cross sections were adjusted to remove anomalies, whilst ensuring the top Castle Reef Formation horizon geometry still adhered to field data. The 3D surface was reconstructed and resampled to make a mesh surface made up of individual triangular segments whose edge lengths measured no more than 20 m. In addition to this 15 more surfaces were created below the top Castle Reef Formation horizon, using lines constructed parallel to the original Castle Reef Formation horizon on each cross section. These arbitrary surfaces were spaced at 10 m intervals, parallel to the top Castle Reef Formation surface above. Field sampling site localities were then projected to the nearest 3D surface, and a value for simple curvature for each sampling site was extracted from the 3D surfaces.

Figure 5 shows the 3D surface for the top Castle Reef Formation horizon, colour mapped for simple curvature, which is the rate of change of dip measured in the direction of maximum dip (e.g. Hennings et al., 2000). The 3D model shows a narrow, high curvature fold forelimb and much lower curvature in the hinge zone and fold backlimb. Curvature in the fold backlimb is consistently low throughout the entire structure, curvature in the hinge zone increases slightly to the SE, and the fold forelimb curvature is much higher in the NW compared to elsewhere along strike. In theory surfaces with higher curvature have undergone more strain to attain those geometries, although in practice this is not always the case (e.g. Chester et al., 1991; Hedlund et al., 1994; Lemiszki et al., 1994; Salvini and Storti 2001; Tavani et al., 2015) so we would predict from our model that fracture intensity should be highest at the northeastern end of the fold forelimb where fold curvature is highest (Figure 5), decreasing along strike. Fracture intensity in the hinge zone and fold backlimb are predicted to be very low in comparison to the fold forelimb due to lower curvatures, but may increase slightly to the south-east, where curvature is slightly elevated. We test these predictions using our field data.

**Results**

Fractures observed on bedding surfaces are predominantly interpreted to be joints, having very few observable offsets. Fractures tend to be very narrow (<1 mm aperture) and are mostly open. The orientations of all measured fractures are presented as a rose plot on Figure 4. In total the orientations of 24744 fractures were measured, inherently producing significant data scatter when presented on
a single rose plot. It is, however, possible to distinguish two dominant fracture set orientations; one being oriented NW-SE, parallel to the fold hinge (i.e. Price’s (1966) J1 fractures); and a second aligning NE-SW, normal to the fold hinge (i.e. Price’s (1966) J2 fractures). These orientation distributions reflect the data from individual sampling sites, where two dominant sets are usually present, aligning normal and parallel to the fold hinge. The orientations of these two fracture sets vary throughout the fold, but may be due to changes in the orientation of the fold hinge along strike (see Figure 4).

Fracture intensity estimations are presented on Figure 4; estimations range from 23 to 464 m/m². From initial observation of the fracture intensity distribution no clear pattern is observed. Estimated fracture intensity is highest in the fold forelimb at the northern end of the structure, where simple curvature is highest (Figure 5); elsewhere fracture intensity distribution appears almost random. To understand the controls on fracture intensity we analyse structural and lithological factors separately.

Fracture intensity: structural controls

Scatter graphs for fracture intensity versus simple curvature and bedding dip for all sampling sites are shown on Figure 6. Due to the nature of exposure on the Swift Reservoir Anticline the majority of our data points are from MSD lithologies (blue, n = 153); with only a small number of sampling sites located on MGSD (yellow, n = 16) and GSD (red, n = 15) bedding surfaces. This means that trends in MSD datasets are likely to be more statistically valid than those from MGSD and GSD datasets. For MSD sampling sites both graphs (Figure 6a and 6b) show positive correlations, with fracture intensity increasing with increasing simple curvature and bedding dip. However both graphs show significant data scatter, resulting in only low-moderate correlation coefficients (R² = 0.41 Figure 6a; R² = 0.30, Figure 6b). For MGSD and GSD datasets Figure 6a and 6b show poor correlation coefficients. Almost no correlation between simple curvature and fracture intensity is observed from either dataset on Figure 6a, and between bedding dip and fracture intensity for MGSD sampling sites on Figure 6b. A weak negative correlation is observed between fracture intensity and bedding dip for GSD sampling sites (Figure 6b). To further investigate how fracture intensity varies at different structural positions, bar charts for the average fracture intensity in forelimb, hinge zone and backlimb positions were plotted for each lithology (Figure 6c). The graph suggests the highest fracture intensities are found in MSD’s outcropping in the fold forelimb, decreasing into the hinge zone and backlimb. Little variation in average fracture intensity in MGSD and GSD lithologies are observed at different structural positions.

To understand the cause behind the data scatter on Figure 6a and 6b we focus on individual sampling transects. Figure 7 shows how estimated fracture intensity varies with simple curvature, bedding dip and structural position for transects 1 and 10 (see Figures 4 & 5 for transect locations). Transect 1 shows that estimated fracture intensity increases with both simple curvature and bedding dip (Figure 7a & 7b). The linear correlation coefficient for both graphs is 0.42, suggesting only moderate data scatter. The cross section (Figure 7c) shows a gradual increase in fracture intensity from the fold backlimb to the fold forelimb.

Scatter graphs for transect 10 show a negative correlation between estimated fracture intensity and simple curvature (Figure 7d) and a positive correlation between estimated fracture intensity and bedding dip (Figure 7e), suggesting a more complicated relationship between structural controls and fracture intensity than indicated by transect 1. The cross section for transect 10 (Figure 7f) shows a gradual decrease in fracture intensity from the fold backlimb to the fold forelimb. A notable difference between transects 1 and 10 is that transect 1 is only sampled in a single lithology (MSD), whereas transect 10 samples three lithologies (MSD, MGSD & GSD); this difference is due to differences in relief and hence stratigraphic position on the two transects. Figures 7d & 7e show that data from each of
the three lithologies are clustered on the graphs, indicative of a lithological control on fracture
intensity.

Fracture intensity: lithological controls

Field lithological classification

Figure 8a shows how fracture intensity varies with lithology. GSD outcrops consistently show low
estimated fracture intensity (min: 27.00 m/m², median: 60.00 m/m², max: 89.00 m/m²), with low
standard deviation (21.97); MGSD outcrops show moderate estimated fracture intensity (min: 58.00
m/m², median: 139.38 m/m², max: 216.25 m/m²), but data points are scattered resulting in a higher
standard deviation (35.30) than for GSD. MSD outcrops exhibit higher estimated fracture intensities
than other lithologies (min: 60.83 m/m², median: 143.75 m/m², max: 463.33 m/m²) but data scatter
also means a high standard deviation (72.52). Average estimated fracture intensity for each of the
three lithologies (Figure 8b) clearly shows a relationship between fracture intensity and lithology;
average fracture intensity is highest in finer grained rocks and decreases as grain size increases. It
should be noted that, since the majority of the Swift Reservoir Anticline exposes the MSD lithology,
average values for fracture intensity have been calculated from a lower number of sampling sites for
MGSD and GSD.

Thin section analysis

The composition of 10 GSD and MSD hand specimens were determined using thin section analysis on
BSE (Back-Scatter Electron) images from an SEM (Scanning Electron Microscope) and image analysis,
using ImageJ software. Examples of three binary images for porosity, calcite and quartz distribution
for the thin section imaged in Figure 9e are shown in Figure 9a-9c. Analysis reveals GSD and MSD hand
specimenss have quite different compositions, textures and porosities. Figure 9d and 9e show typical
BSE images of GSD lithologies; hand specimens are primarily composed of coarse/recrystallized
dolomite (average 90.12 %, Figure 9h) and pore space (average 8.60 %, Figure 9h). Minor components
of quartz (average 1.04 %, Figure 9h), accessory minerals (metal oxides, average 0.05 %, Figure 9h)
and secondary calcite found only in fractures or pore space are also seen (average 0.08 %, Figure 9h).
Figure 9f and 9g show typical BSE images of MSD lithologies; hand specimens are primarily composed
of fine grained dolomite (average 91.60 %, Figure 9h), pore space (average 4.14 %, Figure 9h) and
quartz (average 3.68 %, Figure 9h), with minor components of kaolinite (average 0.28 %, Figure 9h),
accessory minerals (average 0.23 %, Figure 9h) and calcite (average 0.15 %, Figure 9h).

Average percentages for the main mineral components and porosity were calculated for each hand
specimen and plotted against estimated fracture intensity (Figure 10) to investigate the role of
different lithological factors on fracture intensity. A negative correlation ($R^2 = 0.23$) is found between
estimated fracture intensity and porosity (Figure 10a); generally hand specimens with higher
estimated fracture intensity have low estimated porosity values, whereas most hand specimens with
low estimated fracture intensity have higher estimated porosity values. Estimated fracture intensity
shows almost no correlation with dolomite percentage ($R^2 = 0.03$) (Figure 10b), probably because the
dolomite percentage varies very little. A positive correlation between estimated fracture intensity and
quartz percentage is shown on Figure 10c ($R^2 = 0.39$). The graph shows that hand specimens with
higher quartz content generally have higher fracture intensity than those with lower quartz content.
A positive correlation is also seen between estimated fracture intensity and kaolinite content ($R^2 =
0.39$, Figure 10d). Hand specimens with low kaolinite content tend to have low estimated fracture
intensity, whereas hand specimens with more kaolinite have higher estimated fracture intensities.

Figure 10e shows a moderate correlation ($R^2 = 0.15$) between estimated fracture intensity and calcite
percentage; however given that only three of our hand specimens contain calcite, and that most of that calcite is observed in fractures/pores (Figure 9) suggests that it was probably not present during the main phase of fracturing, and therefore calcite is unlikely to have influenced fracture intensity. A positive correlation is also seen between quartz and kaolinite content (Figure 10f); the implications of this correlation will be considered in the discussion.

Discussion

Structural versus lithological control on fracture intensity

Correlation between fold simple curvature, bedding dip, and structural position with estimated fracture intensity can be seen in our data (Figure 6). Estimated fracture intensity increases as simple curvature and bedding dip increase, suggesting that these two factors are related to fracture formation. These relationships are clearest where data is collected from a single lithology; for example data from Transect 1 (Figure 7) is collected in only Mud-Supported Dolomites. Here positive linear correlation coefficients for these variables are moderate (Figure 7a-c). Generally higher simple curvatures and bedding dips are found in the fold forelimb, where fracture intensity is highest, and they decrease south-westward into the fold backlimb, where fracture intensities are lowest (Figure 7a-c). When analysing the data in more detail we see that these relationships between structural controls and estimated fracture intensity only hold true where lithology is consistent (e.g. Figure 7d-f); our data shows that, regardless of structural position, fracture intensity will be higher in mud-supported dolomites than grain-supported dolomites. This means that the high curvature, steeply dipping fold forelimb, if sampled in a mud-supported dolomite fracture intensity will be high, whereas in a grain-supported dolomite, fracture intensity will be lower (Figure 11).

Relationships between fold curvature and fracture intensity have been discussed at length in published literature, and stem from work by Ramsay (1967), who proposed relationships between strain and fold curvature in two dimensions. Lisle (1992; 1994) further developed work by Ramsay (1967) by assessing how strain and curvature relate in three dimensions. Lisle uses curvature analysis to detect zones high strain on folded surfaces, and suggests that this curvature analysis could be used to predict the density of sub-seismic scale deformation such as fracturing (Lisle, 1994). Lisle (1994) uses the theory that surfaces with double curvature (i.e. a non-cylindrical fold) must form with some stretching or contraction of the bedding, meaning that the total curvature (product of the two principle curvatures) is proportional to strain magnitude. This theory was tested by Lisle (1994) on the Goose Egg Dome in Wyoming, where the highest fracture densities, measured by Harris et al., 1960, were found in regions of highest total curvature. Other studies have used this predicted relationship between curvature and strain/fracture intensity to predict fracture intensity on fold structures. Hennings et al. (2000) use the rate of change of dip (i.e. simple curvature) on the Oil Mountain Anticline in Wyoming to populate a 3D fold model with fracture intensity estimates based on the assumption that fracture intensity increases with increasing curvature.

However, in many other investigations into the role of curvature and structural position in controlling fracture intensity and density, it has been found that correlations are poor. Examples include Ortega et al. (2010), whose suggest that only a moderate to weak correlation between fracture intensity and fold curvature can be observed in carbonates of the Sierra Madre Oriental in northeast Mexico. McQuillan (1973) uses outcrops from the Asmari Formation in the Zagros to determine that fracture density is independent of structural position. Bergbauer & Pollard (2004) suggest that fracture intensity is higher on sandstone folds than in between fold structures, however there is little variation...
in fracture intensity within individual fold structures. Our study suggests that although fracture intensity is influenced by structural factors such as bedding dip, fold curvature and structural position, it is the lithology that is the main control on fracture intensity distribution across the fold structure. The implications for this are that fracture intensity prediction in subsurface structures may be better aided by determining how lithology changes spatially rather than focussing on characterising the geometry of a fold and modelling its evolution.

Based on fracture orientations we classified our fracture sets as Price’s (1966) hinge-parallel (J1) or hinge-normal (J2) fractures that form in response to folding. An alternative model could be that the fractures observed are in fact pre-folding joints associated with foreland flexuring in the peripheral bulge region that are later tilted as the fold develops (Tavani et al., 2015). The elevation of fracture intensity in the forelimb could then be caused by late-stage fold tightening that increases fracture frequency in this region alone (i.e. backlimb and hinge-zone fractures are all pre-folding features). If this is the case then folding-associated strains influence fracturing in the forelimb region alone, not the backlimb and hinge zone regions. The resultant fracture sets for both models (pre-folding and fold-associated fracturing) are likely to produce very similar fracture patterns (i.e. hinge-parallel and hinge-normal joints perpendicular to bedding), so it is difficult to distinguish between them.

**Lithological control on fracture intensity**

Using image analysis of Back-Scatter Electron (BSE) photographs we were able to determine that fracture intensity shows some correlation with the proportion of porosity, quartz and kaolinite (Figure 10). The reason for this probably relates to these lithological factors influencing overall rock strength. It is thought that fracture intensity increases with increasing rock strength because stronger rocks tend to be more brittle, so that when they fail they produce closely spaced, high density fracture networks (Nelson, 2001). Our data suggests that fracture intensity increases with decreasing porosity (Figure 10a); similar correlations have been observed elsewhere, for example Nelson (2001) reports increasing fracture intensity with decreasing porosity in carbonates further south in the Sawtooth Range. The negative correlation between fracture intensity and porosity observed in our study could be explained by strain accommodation during deformation. One explanation might be that pore spaces act as weak zones in the rock; as stress is applied to the rocks during folding, the pore space may accommodate strain by distortion of the pore boundaries (i.e. elastic behaviour). Bounding grains could be pushed into the pore space, accommodating a significant portion of the overall strain, meaning only a limited number of fractures need to form to accommodate the remaining strain. In rocks with very low porosity, such as the mud-supported dolomites in our study area, limited pore space might mean very little pre-failure strain can be accommodated by pore space rearrangement (elasticity) so instead a large number of fractures form. If this was the case we might expect to see evidence for pore shape deformation. Although our BSE images of thin sections clearly show many pores, we cannot assess the degree of pore shape deformation because we do not have any evidence of the pore shapes prior to folding.

Compositional correlations with estimated fracture intensity are also observed from our data. Fracture intensity increases with both increasing quartz and kaolinite content (Figure 10c, 10d). Quartz is a strong, brittle mineral; increasing the amount of quartz in a rock will probably also increase the bulk strength and brittleness of that rock. Based on Nelson’s (2001) suggestion that rocks with a higher percentage of brittle constituents will have closer spaced (and therefore higher intensity) fractures, it would be expected that rocks with more quartz will have a higher fracture intensity than those with
less brittle constituents, which fits our data. The positive correlation between estimated fracture intensity and kaolinite percentage is more puzzling. Kaolinite is a clay mineral and as such is usually considered weak and incompetent in comparison to other minerals such as dolomite and quartz; rocks with more weak and incompetent minerals would have lower bulk strengths and be less brittle than those without, therefore we might expect lower fracture intensities in rocks containing more clay. This relationship is observed elsewhere; Corbett et al. (1987) determine that chalks containing smectite are weaker than those without because large clay masses act as soft inclusions that concentrate the applied stress. Ferrill and Morris (2008) also suggest that clay rich carbonates are incompetent so they are able to accommodate more pre-failure strain, resulting in lower fracture intensities at outcrop. Both of these studies suggest the opposite relationship to what we observe from our data; we suggest our observed positive correlation between fracture intensity and kaolinite content probably relates to the fact that kaolinite-bearing rocks also tend to have significant proportions of quartz (Figure 10f), which controls the bulk rock properties and resulting fracture intensity. Kaolinite percentages are very low (<0.7 %) so probably do not have a significant impact on bulk rock properties; the correlation between kaolinite content and estimated fracture intensity may be coincidental.

Our data suggests porosity and compositional factors are the main lithological controls on fracture intensity, however there are several other studies that suggest other lithological properties influence fracture intensity. Mechanical layer thickness is thought to correlate with fracture intensity; it is thought that thicker beds will have wider spaced (and therefore lower intensity) fractures. Studies find this relationship in carbonates (McQuillan, 1973; Huang & Angelier, 1989; Wennberg et al., 2006; Wennberg et al., 2007; Barbier et al., 2012) and siliciclastics (Hobbs, 1967; Gross et al., 1995; Florez-Niño et al., 2005). A limitation of our study area is that generally only the top surfaces of beds are exposed so we cannot measure mechanical layer thickness to test this hypothesis. Another lithological property that we have not been able to test is grain size, which has been seen to correlate with fracture intensity in other field examples. Although we have backscatter electron images from our hand specimens they only clearly show quartz grain boundaries; dolomite grain boundaries are very difficult to identify, therefore we could not calculate average grain sizes for our hand specimens. Hanks et al., 1997, suggest coarser grained rocks contain lower fracture intensities in carbonates. This would fit our qualitative observations that lower fracture intensities are found in coarser grain-supported dolomites on the Swift Reservoir Anticline. However, as Nelson (2001) points out, although a increasing fracture intensity is often attributed to decreasing grain size, thinner beds often have lower grain sizes so correlations between fracture intensity and grain size may actually relate to the mechanical layer thickness.

**Conclusions**

Based on our data we suggest that the greatest controls on fracture intensity are porosity and quartz content, followed by structural factors such as simple curvature and structural position. Where lithology is constant (i.e. sampling at the same stratigraphic position along a transect) fracture intensity has a positive correlation with fold simple curvature and bedding dip; the highest fracture intensities are found in fold forelimb outcrops. Where variations in lithology occur (i.e. sampling multiple bedding surfaces at different stratigraphic positions in the Castle Reef Formation), the fracture intensity is unpredictable based on fold geometry; instead the fracture intensity is at least partially controlled by porosity and quartz percentages that control the bulk rock strength and its mechanical behaviour under stress. The implications of these results are that regions of a fold that have undergone high stresses during folding, such as high curvature hinge zones and forelimbs that may be preferentially be targeted during hydrocarbon exploration, may not necessarily provide well
connected fracture networks if the lithology is porous and quartz-poor. In this situation a better fracture network may be found elsewhere in a region of lower curvature but higher quartz content and lower porosity. Although structural factors influence fracture formation, it is the mechanical properties of the rock that are the main control on fractured reservoir quality.

Acknowledgements

This research was funded by Oil Search Ltd, Santos Ltd and InterOil, through the University of Aberdeen Fold-Thrust Research Group. Electron Microscopy was performed in the ACEMAC Facility at the University of Aberdeen with assistance from John Still. Joyce Neilson is thanked for advice on the use of ImageJ software. Midland Valley are thanked for the use of their Move software for field data collection and model building. We thank Alfred Lacazette and Stefano Tavani for reviewing the manuscript and providing constructive comments.

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Figure captions

Figure 1: Simplified geological map of the Sawtooth Range and surrounding area. Extent of Figure 3a is shown by the black box. Adapted from Mudge et al., 1982; Mudge & Earhart, 1983.

Figure 2: a) Sedimentary log through the fold forelimb on the southern side of the Swift Reservoir dam cut. Bed thickness here decreases with decreasing grain size. b) Field photograph of the southern side of the dam cut showing the location of the sedimentary log (Figure 2a). Three lithological units are identified: mudstone, wackestone-packstone, grainstone. The majority of fracture sampling sites are located higher in the stratigraphy than the log, for example on the mudstone outcrops at the top of the image. c) An example of a fractured mudstone (MSD) bedding surface. d) An example of a fractured grainstone (GSD) bedding surface.

Figure 3: a) Geological map of Swift Reservoir and surrounding region showing the location of the Swift Reservoir Anticline (SRA) and line of section (Figure 3d) adapted from Mudge & Earhart, 1983. b) 3D photogrammetric model of the Swift Reservoir Anticline showing its asymmetric structure. c) Key to geological map and cross section (Figure 3a, Figure 3d). d) Cross section through the Swift Reservoir Anticline showing the local structural style. MSBT, Major Steel Backbone Thrust; ST, Swift Thrust; SRA, Swift Reservoir Anticline; ECIZ, Eagle Creek Imbricate Zone; IZ, Eagle Creek Imbricate Zone in the subsurface; FLT, Fish Lake Thrust; OMT, Old Man Thrust; MLT, Mitten Lake Thrust; BT12-1, Blackfeet Tribal 12-1 well; W-1, Wickware 1. e) Schematic restoration of the Swift Reservoir Anticline showing...
an asymmetric anticline with a steep forelimb that could result from fault propagation folding or trishear or the combination of both. The solid red line shows the extent of the thrust at this development stage and the dashed red line shows where the thrust will subsequently cut through the fold to emplace the fold to its current location.

Figure 4: Aerial photograph (from Google Earth) showing the locations of sampling sites and cross sections (T1-T11) used to construct the 3D model (Figure 5). The size of the circles at each sampling site reflect the magnitude of fracture intensity, and the colour of the circle reflects the lithology at each sampling site. Transect/cross section locations are shown by orange lines; the fold hinge is shown by the green line; structural zone (forelimb, hinge zone, backlimb) boundaries are shown by yellow lines.

Figure 5: 3D model of the top Castle Reef Formation constructed from parallel cross sections colour-mapped for simple curvature. The positions of cross sections/field transects (T1-T11) are shown as orange lines.

Figure 6: a) scatter graph of fracture intensity versus fold simple curvature (extracted from the 3D model-Figure 5), showing a positive correlation for MSD sites, and almost no correlation for MGSD and GSD sites. b) scatter graph of fracture intensity versus bedding dip, showing a positive correlation for MSD sites and negative correlations for MGSD and GSD sites. c) Bar chart showing the average fracture intensity for MSD, MGSD and GSD sampling sites in different structural positions. For all graphs blue datapoints are from Mud-Supported Dolomites (MSD), yellow are from an intermediate phase (MGSD) and red are from Grain-Supported Dolomites (GSD).

Figure 7: a) Scatter graph of fracture intensity versus fold simple curvature (extracted from the 3D model-Figure 5) for transect 1 sampling sites. b) Scatter graph of fracture intensity versus bedding dip for transect 1 sampling sites. c) Cross section through transect 1 showing the change in fracture intensity with structural position. d) Scatter graph of fracture intensity versus fold simple curvature (extracted from the 3D model-Figure 5) for transect 10 sampling sites. e) Scatter graph of fracture intensity versus bedding dip for transect 10 sampling sites. f) Cross section through transect 10 showing the change in fracture intensity with structural position. g) Key to cross sections. See Figures 4 & 5 for transect locations. For all figures blue data points are from Mud-Supported Dolomites (MSD), yellow are from an intermediate phase (MGSD) and red are from Grain-Supported Dolomites (GSD).

Figure 8: a) Scatter graph showing fracture intensity distribution in GSD, MGSD and MSD lithologies. Box and whisker plots for each lithology show the minimum, maximum, first quartile, third quartile and median. b) Bar chart showing the average fracture intensity for each lithology observed in the field. Fracture intensity decreases from mud-supported to grain-supported dolomites.

Figure 9: a) Binary image showing the porosity distribution (black) for Figure 9e; porosity for this image is 32.51%. b) Binary image showing the calcite distribution (black) for Figure 9e; calcite percentage for this image is 1.90%. c) Binary image showing the quartz distribution (black) for Figure 9e; quartz percentage for this image is 1.51%. d-g) Representative BSE (Back-Scatter Electron) images showing GSD (d-e) and MSD (f-g) lithologies. h) Average composition for GSD (red) and MSD (blue) hand specimens.

Figure 10: scatter graphs for fracture intensity versus a) porosity percentage, b) dolomite percentage, c) quartz percentage, d) kaolinite percentage, e) calcite percentage. f) Scatter graph for quartz percentage versus kaolinite percentage. Moderate correlations are seen between fracture intensity and porosity, quartz content and kaolinite content. Fracture intensity has a weak correlation with dolomite content and calcite content. A positive correlation between quartz percentage and kaolinite percentage is observed.
Figure 11: Schematic model of fracture intensity variation on the Swift Reservoir Anticline: fracture intensity increases with increasing fold curvature in a given lithology. Fracture intensity is much lower in grain-supported dolomites with higher percentages of quartz and porosity than mud-supported dolomites, regardless of fold curvature and structural position.
FIGURE 2

(a) Stratigraphic column showing mudstone, wackestone-packstone, grainstone, chert nodules, slip horizon, planar bedding, fossil frag., and structureless.

(b) Depiction of bedding surface at 100° and 135°.

(c) Close-up of mudstone with scale 20 cm.

(d) Close-up of grainstone with scale 20 cm.
FIGURE 3

- Upper Cretaceous (Two Medicine Fm.)
- Mississippian (Castle Reef Fm.)
- Thrust fault
- Horizon boundary
- Plunging anticline hinge
- Bedding strike/dip
- River/lake
- Well
- Upper Cretaceous (Marias River Shale)
- Upper Cretaceous (Virgelle Sandstone & Telegraph Creek Fm.)
- Lower Cretaceous (Mount Pablo Fm.) & Jurassic (Morrison Fm. & Ellis Group)
- Mississippian (Castle Reef Fm.)
- Mississippian (Allan Mountain Fm.)
- Lower Cretaceous (Blackleaf Fm.)
- Lower Cretaceous (Kootenai Fm.)
- Devonian (Three Forks, Jefferson & Maywood Fm’s)
- Cambrian

- Asymmetric fold structure
- Fractured bedding surfaces
- Swift Reservoir Dam
- NE-SW direction
- 5000 m scale
- SRA
- BT
- MSBT
- ST
- ECIZ
- FLT
- MLT
- Thrust A
- 12-1
- 12
- OMT
- SW-NE direction
- 5000 m scale
FIGURE 4

- Lithology
  - MSD (Mud Supported Dolomite)
  - MGSD (intermediate phase)
  - GSD (Grain Supported Dolomite)
  - unspecified lithology

- Fracture intensity (m/m²)
  - 0-99
  - 100-199
  - 200-299
  - 300-399
  - 400-499

- Fold hinge
- Plunge
- Cross sections
- Forelimb
- Backlimb
- Hand specimens

n = 24744
FIGURE 5
FIGURE 6
Transect 1

(a) Simple curvature vs. Fracture intensity (m/m²)

(b) Bedding dip (°) vs. Fracture intensity (m/m²)

(c) SW - NE transect with marker horizons

Transect 10

(d) Simple curvature vs. Fracture intensity (m/m²)

(e) Bedding dip (°) vs. Fracture intensity (m/m²)

(f) SW - NE transect with marker horizons

(g) Lithological classification and Fracture intensity (m/m²)

- MSD
- MGSD
- GSD
- Unclassified

FIGURE 7
FIGURE 8
FIGURE 9

Porosity
Quartz
Calcite
Kaolinite
Other
Dolomite

average % (porosity, quartz, calcite, kaolinite & other)

Grain supported dolomite (GSD)
Mud supported dolomite (MSD)

average % (dolomite)

(h)
FIGURE 10

(a) Porosity (%) vs. Fracture intensity (m/m²)
- $R^2 = 0.23$
- $n = 10$

(b) Dolomite (%) vs. Fracture intensity (m/m²)
- $R^2 = 0.03$
- $n = 10$

(c) Quartz (%) vs. Fracture intensity (m/m²)
- $R^2 = 0.39$
- $n = 10$

(d) Kaolinite (%) vs. Fracture intensity (m/m²)
- $R^2 = 0.39$
- $n = 10$

(e) Calcite (%) vs. Fracture intensity (m/m²)
- $R^2 = 0.15$
- $n = 10$

(f) Kaolinite (%) vs. Quartz (%)
- $R^2 = 0.60$

Grain supported dolomite (GSD) - Red circle
Mud supported dolomite (MSD) - Blue circle