Fracture abundance and strain in folded Cardium Formation, Red Deer River anticline, Alberta Foothills, Canada

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ABSTRACT

The folded and thrust Mesozoic clastic sequence of the Canadian Rocky Mountain foothills forms important hydrocarbon reservoirs; however, the generation of fractures and their timing relative to folding and thrusting have remained unclear. Understanding the distribution and evolution of natural fractures, especially their timing of formation relative to the evolution of the fold-and-thrust system, could potentially improve exploration and development outcomes in these otherwise tight, low-permeability reservoirs. We investigated the relation between folding and fracture formation in the Upper Cretaceous Cardium Formation by combining field structural observations and kinematic modeling of the fold-and-thrust belt evolution. We evaluated the effects of structural position and strain on the formation of predominantly opening-mode fractures by analyzing fracture intensity and aperture in the backlimb, the shallow forelimb close to the crest, and the steep forelimb, away from the crest of a reservoir-scale anticline, the Red Deer River anticline. Fracture strain determined from both outcrop and microscanlines varies little across the three structural domains of the fold, although somewhat lower fracture strain was measured in the forelimb close to the crest. These fracture strain measurements are qualitatively consistent with calculated horizontal strain in the tectonic-transport direction obtained from kinematic numerical models that simulate fold development associated with slip along the underlying Burnt Timber thrust. The kinematic models predict similar magnitudes of horizontal extension in both the back and forelimbs, and somewhat lower extension in the upper forelimb during early development of the Red Deer River anticline. Predicted early fracture formation during fold development is consistent with field structural observations of shear reactivation during later stages of folding. This combined kinematic modeling and field structural study demonstrates that deforming fold and thrust belts can undergo a complex evolution of bed-parallel extension in both space and time, resulting in spatially variable fracture formation in structurally complex subsurface reservoirs.

Keywords: Cardium; Alberta Foothills; Fracture abundance; Strain

INTRODUCTION

The Canadian Rocky Mountains are one of the most extensively studied foreland fold-and-thrust belts, with specific focus on the Foothills and Front Ranges because of their hydrocarbon potential (Leckie and Smith, 1992; Price, 1994; Boettcher et al., 2010). The Foothills of the Canadian Rocky Mountains have been explored for oil and gas since the early 20th century. Initially, exploration primarily focused on what are now termed conventional hydrocarbon reservoirs, but the low-permeability hydrocarbon reservoirs have been the focus in recent years. Since natural fractures affect permeability, and thus production from these tight reservoirs, an understanding of the spatial distribution of natural fractures, their structural and diagenetic attributes, and the spatial and temporal relationships of their formation relative to the fold-and-thrust belt evolution are important (Currie and Nwachukwu, 1974; Currie and Reik, 1977; Narr and Currie, 1982; Newson, 2001; Salvini and Storti, 2001; Jager, 2008). However, the spatial and temporal relationships between fracture formation and the development of folds and thrusts based on field structural criteria are frequently ambiguous. For example, folds in fold-and-thrust belts may undergo a complex evolution, with regions of highest curvature and folding-induced strain migrating in space and time (e.g., Salvini and Storti, 2004), making it difficult to reconstruct the conditions under which fractures formed relative to the fold geometry observed in outcrop or seismic profiles.

Production rates in the Cardium Formation are regionally variable (e.g., Van and Pedersen, 2015), which may be related, at least in part, to regional differences in natural fracture characteristics. Moreover, well tests, flow rates, and production flow surveys that are incompatible with production from the rock matrix alone have convinced operators in western Canada that natural fractures contribute to production (e.g., Cooper, 1992; Jamison, 1997;

Newson, 2001; Solano et al., 2011). However, uncertainty about where and when fractures form relative to the fold-and-thrust belt regional structure means that the location of abundant, open fractures cannot be accurately predicted. This leads to unacceptable risk in exploration and development decisions (Nelson, 1985; Newson, 2001)

Formation of fractures related to folding was conceptualized several decades ago (Price 1966; Stearns, 1968; Friedman, 1969; Stearns and Friedman, 1972; Hancock, 1985; Price and Cosgrove, 1990), and many recent studies have inferred fracture formation related to folding in fold-and-thrust belts (Hennings et al., 2000; Hanks et al., 2004; Florez-Niño et al., 2005; Bellahsen et al., 2006; Tavani et al., 2006, 2008; Ghosh and Mitra, 2009; Smart et al., 2012). However, fractures can also predate folding and potentially become reactivated in opening-mode or by shearing during folding (Dunne 1986; Laubach, 1988; Bergbauer and Pollard, 2004). Reactivation also leads to formation of new fractures (Davatzes and Aydin, 2003). Furthermore, fractures may form during uplift and exhumation concurrently with, or after, folding and thrusting (Currie and Reik, 1977; Engelder, 1985; Casini et al., 2011; English, 2012; Reif et al., 2012; Fall et al., 2012, 2015). Fractures formed during later stages of exhumation may not be of significance to deep petroleum reservoirs.

Interpretation of fracturing associated with folding may be hindered by lack of direct observation of the changes in fracture patterns through time within an evolving fold-and-thrust sequence unless fracture cementation is syntectonic – during deformation- and all deformational stages are recorded. Previous studies that addressed the spatial and temporal relationships between fracture formation and fold-and-thrust development were based on 1) field observations of cross-cutting relationships between fractures, folds, and associated structures such as stylolites (Muecke and Charlesworth, 1966; Nickelsen, 1979; Barton, 1983; Graham et al., 2003), 2)

comparison of numerically modeled folds and natural folds with syn-tectonic growth strata (Verges et al., 1996; Ford et al., 1997; Poblet et al., 1997; Novoa et al., 2000; Salvini and Storti, 2002; Shackleton et al., 2011), 3) structural restoration algorithms (Rouby et al., 2000; Griffiths et al., 2002; Thibert et al., 2005; Maerten and Maerten, 2006), and 4) forward geometric and geomechanical fold models (Erslev, 1991; Johnson and Johnson, 2002; Poblet and McClay, 1996; Suppe, 1983; Suppe and Medwedeff, 1990; Smart et al., 2012). We combine field-based observations of cross-cutting relationships with forward kinematic models to show that fractures either predate or formed at the onset of folding, and subsequently evolved during continued thrusting and folding.

In this study, we document relative timing and spatial distribution of fractures in a reservoir-scale fold, the Red Deer River anticline, Alberta, Canada (**Figure 1**). Fractures are observed in the Cretaceous Cardium Formation, a sandstone with low porosity and permeability that is a significant reservoir in the Alberta Foothills. By comparing observed fracture distributions to the modeled evolution of this fault-related fold, we test whether fractures are kinematically related to folding or pre- or postdate folding. This study combines (i) field investigations to analyze fracture characteristics such as aperture and cement fill within a structural context, (ii) macro- and microscanline analyses to measure fracture abundance and strain within structural position, and (iii) kinematic models to both constrain the timing of fracturing in the Foothills relative to the overall structural evolution of the fold-and-thrust belt in this area, and to compare strain distributions from the models to those measured using scanlines in the field. If fractures formed concurrently with folding, fracture abundance is expected to correlate with folding-associated strain. Alternatively, if fractures formed prior to or post-folding, no correlation between fractures and folding strain is expected. A comparison between

field fracture strain data and strain calculated by numerical fault and fold kinematic models indicates that 1) opening-mode fractures within the Cardium Formation in the Red Deer River anticline either predate folding or formed during the early onset of folding, and 2) cross-fold, opening-mode fractures were reactivated in shear and linked through propagation during folding.

GEOLOGIC SETTING

The Red Deer River anticline is located in the central portion of the Alberta Foothills that form the eastern part of the Rocky Mountains, a thin-skinned, Late Jurassic to Eocene foreland fold-and-thrust belt (Monger, 1989) (**Figure 1**). The structural framework of the Alberta Foothills contains numerous steeply-dipping, northeasterly-verging imbricate thrust sheets of the Mesozoic foreland basin and eastward-tapering sedimentary wedge rocks that were deposited on the North American crystalline basement (Price, 1994; Fermor, 1999). During the Paleozoic era, marine carbonates, shales, and sandstones were deposited along the western margin of North America. In contrast, lower Jurassic to Paleogene rocks comprise sandstone, siltstone, conglomerate, and shales that were deposited in the foreland basin during the rise of the Canadian Cordillera (Bally et al., 1966; Price, 1973; Beaumont, 1981; Stockmal et al., 1992). The deformation history in the Canadian Rocky Mountains is characterized by northeastward shortening during the Laramide orogeny from Campanian to Eocene time (Price, 1994, Eisbacher et al., 1974; van der Pluijm et al., 2006), although there is evidence that deformation east of the Rocky Mountain Trench began as early as Late Jurassic (e.g., Pana and van der Pluijm, 2015). The mechanical contrast between Paleozoic and Mesozoic rocks exerts a strong control on the deformation style of the orogen (Price, 1981; Taerum, 2011).

The clastic succession of Mesozoic rocks includes the prominent Upper Cretaceous Alberta Group. This group consists of predominantly marine to coastal marine foreland basin

fills (Stott, 1963) (**Figure 2**). A prominent sandstone unit of this group is the Cardium Formation, which separates thick sections of shale of the Wapiabi Formation above and the Blackstone Formation below the Cardium . It has been proposed that these thick shales mechanically decouple the Cardium Formation from deformation in the overlying and underlying competent units (Wall, 1967; Taerum, 2011). The Cardium Formation is approximately 150 m thick where exposed in the central Alberta Foothills, and highly variable but around 50 m thick in the subsurface under the Plains to the east (Krause et al., 1994). The Cardium Formation contains large hydrocarbon reserves east of the Foothills, where it is forming the youngest oil-bearing formation and it is also a major producer in the Pembina field southwest of Edmonton (Krause et al., 1994; Newson, 2001).

Cardium Formation of the Red Deer River Anticline

We investigated fracture distribution within the Cardium Formation exposed on the slopes of the Red Deer River valley across from the Mountain Aire Lodge west of Sundre, Alberta, Canada (**Figure 3**). The Cardium Formation is exposed for about 1 km along a section oriented approximately perpendicular to the axis of the NW-plunging Red Deer River anticline (Olleranshaw, 1996; Dechesne and Veilleux, 2000). Bedding dip varies from approximately 25-30° W in the western limb to 35-40° E in the eastern limb, and average approximately 50° E in the easternmost part of the east limb. The strata forming the crest of the anticline are not exposed. The Cardium Formation in the study area is composed of very fine to medium-grained, gray to dark gray-brown sandstone layers. The sandstones exhibit differing types of internal lamination, including hummocky cross-stratification and varying amounts of bioturbation. In the Red Deer River anticline, the Cardium Formation can be subdivided into three members from the base to the top: the Kakwa, Low Water, and Karr members, each of which is separated by finer-grained,

slope-forming mudstone and siltstone units (Dennis Meloche, personal communication, 2012). The classification of members at this locality is based upon similarity in stratigraphic position relative to the Musreau/Kakwa delta succession to the northwest (Plint et al. 1988). The three members of the Cardium Formation are well-exposed on the west and east limbs of the Red Deer River anticline on the hillside, and in two smaller outcrops (**Figure 3**) located on the north and south banks of the Red Deer River adjacent to the Mountain Aire Lodge. We divided the Red Deer River anticline into three structural domains. The west limb of the anticline is hereafter referred to as Domain I, and the east limb of the anticline is divided into two domains; Domain II spans the shallow part of the east limb near the eroded hinge, and Domain III spans the more steeply dipping part of the east limb (**Figure 3a**).

The Kakwa Member is most completely exposed on the western limb of the fold (Domain I) and it consists of well-developed upper and lower parts. Both horizons are thinner –or less completely exposed- in the east limb of the fold (Domain II) to the point that the lower portion of the Kakwa is not visible from a distance. The overlying Low Water Member displays approximately constant thickness along the 1 km-wide exposure, and is both finer-grained and more highly bioturbated than the Kakwa Member. The uppermost Karr Member shows the greatest variability; on the west limb it is composed of thinner, amalgamated, hummocky cross-stratified, upper flow regime sandstone incised into thicker bioturbated, fine-grained muddy silty turbidites; on the east limb it is composed of very fine- to fine sandstone with planar to hummocky cross-stratified bed forms. Domain III comprises the most steeply-dipping part of the east limb, which outcrops on the hillside and along a small exposure located on the south side of the Red Deer River (**Figure 3f-g**). This small exposure is inferred to be part of the Kakwa Member based on lithological properties and stratigraphic/structural location.

All samples analyzed are fine-grained litharenites containing on average 15-20% chert and 10-15% mud clasts. Bioturbated litharenites of the Kakwa Member comprise up to 30% mud clasts. Monocrystalline quartz grains are more sutured within the planar-laminated and cross-stratified Kakwa and Karr sandstones, versus the bioturbated Kakwa sandstone. Karr litharenites contain up to 5% intergranular porosity in these exposures, whereas porosity in the Kakwa sandstones is nearly zero.

FRACTURE CHARACTERIZATION

Methods

To quantify the extensional strain accommodated by opening-mode fractures in the direction of tectonic transport, we measured fracture aperture and spacing within the Cardium Formation in the direction perpendicular to the trend of the Red Deer River anticline along one dimensional scanlines in the field and in thin section. Because only extension parallel to the tectonic transport direction was modeled in this study, one dimensional scanlines oriented parallel to the transport direction, measuring the strain tensor component parallel to the one dimensional scanline, was considered adequate. To avoid inclusion of fractures that formed during late-stage exhumation and thus potentially unrelated to folding, we measured spacing and aperture only of completely cemented or cement-lined fractures. We will demonstrate below that fracture cements formed during fracture opening under deep burial conditions unlike barren fractures that are assumed to form during late-stage exhumation. Barren fractures were not included in the scanlines.

For each fracture measured in outcrop, we recorded kinematic aperture (distance orthogonal to the fracture from one fracture wall to the opposite wall, counting cement infill and remaining fracture porosity as fracture aperture, fracture spacing, and cement infill type).

Fracture aperture was measured using a hand lens against an aperture comparator as described by Ortega et al. (2006). We refer to the scanlines measured in the field as macroscanlines, and fractures ≥ 0.1 mm in aperture as macrofractures.

Stress release during exhumation can widen fractures that originally formed at reservoir depth (Engelder and Plumb, 1984; Meglis et al., 1991). Based on fluid inclusion data presented below we infer that fractures in this area formed at reservoir conditions. If reactivated, fracture cements containing bridging cement would break. Therefore, completely cemented fractures, or fractures containing intact cement bridges provide fracture apertures not affected by stress release during exhumation. For cement-lined fractures or cemented fractures that are split open we measured the cement thickness as a minimum estimate of kinematic aperture. We collected two macroscanline datasets; one from the Karr Member in the back (west) limb (Domain I), and one from the upper portion of the Kakwa Member within the gently dipping fore (east) limb (Domain II) of the fold (Figure 3). Macroscanlines were collected from outcrops with minimal surface weathering and continuous exposure of sufficient quality that all cemented or cement-lined fractures ≥ 0.1 mm in aperture along the transect could be reliably identified.

To quantify fracture strain accommodated by fractures of apertures < 0.1 mm (microfractures) we collected three field samples with representative fracture populations from within the two macroscanlines for laboratory microfracture analyses: two samples from scanline 1 in Domain II, and one sample from Scanline 2 in Domain I. Eight additional field samples were collected for laboratory microscanline analysis at four other locations across the fold where outcrop quality did not allow the collection of macroscanlines. To account for variability due to lithological differences across the Cardium Formation, three of these additional field samples were collected from three lithologically different layers within the Kakwa Member in the

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4 245 steepest part of the forelimb (Domain III), and two samples were collected from lower and upper
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6 246 portions of the Kakwa Member on the back limb near the crest of the anticline (Domain I).
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9 247 Fracture strain calculations using macro- and microfracture data are explained below.
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12 248 In sum, fracture scaling analyses were performed on two macroscanlines and 38
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14 249 microscanlines from eleven field samples.
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17 250 For microscanlines, we prepared contiguous polished thin sections from field samples
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19 251 following the method described by Gomez and Laubach (2006). Microscanline length ranged
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21 252 from 2.6 to 25.8 cm. Thin sections were imaged with an Oxford MonoCL cathodoluminescence
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23 253 (CL) system attached to a Philips XL30 SEM at the Bureau of Economic Geology, The
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25 254 University of Texas at Austin. Operating conditions were at 15 kV using a blue filter, and CL
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27 255 images were assembled into contiguous mosaics following the protocol described by Gomez and
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29 256 Laubach (2006) and Hooker et al. (2015). CL images provide information about fractures that are
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31 257 not discernible with the naked eye or using transmitted light petrography (Milliken and Laubach,
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33 258 2000). CL images were imported into Didger 3 digitizing software where fractures were spatially
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35 259 referenced. Spatial X-Y coordinates generated by Didger 3 were imported into an Excel routine
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37 260 that allows for the calculation of fracture aperture, spacing, and other fracture statistics based on
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39 261 spatial coordinates (Gomez and Laubach, 2006). Only transgranular fractures were selected, and
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41 262 fractures were then subdivided into sets based on their orientation and systematic cross-cutting
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43 263 relationships.
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52 264 **Fracture orientations**

53 54 55 265 Opening-mode fractures 56 57

58 266 We distinguish four sets of bed-perpendicular, opening-mode fractures based on
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orientation and cross-cutting relationships (sets 1-4), and one bed-parallel set (set 5) (**Table 1**). All bed-perpendicular but set 4 are bed-bounded, i.e., fractures terminate against bedding surfaces or are confined within a bed (**Table 1**). In the classification of Hooker et al. (2013) the height pattern is top-bounded on the scale of individual beds, but with a hierarchical pattern of more widely spaced fractures that cut several beds. Set 1 fractures are sub-perpendicular to the fold axis (cross-fold), set 2 fractures are oblique to the fold axis (cross-strike), and set 3 are subparallel to the fold axis (strike-parallel) (**Figure 4**). In Domain I, set 1 fractures strike roughly N-S, set 2 fractures strike NW-SE, and set 3 fractures strike E-W (**Figure 5a**). In Domains II and III, the relative orientations are preserved, but all fracture sets rotate progressively clockwise by 40-60° consistent with the change in bedding across the fold. When bedding is rotated to horizontal, fractures of all bed-perpendicular sets show similar orientations across the different structural domains of the fold (**Figure 5b**).

Set 4 fractures, several meters tall, cut the sandstone layers of the Kakwa Member from top to bottom (**Figure 6a, c**), and are best developed within Domain II. Although of the same orientation as set 1 fractures, these through-going set 4 fractures are interpreted as a separate set because they are closely spaced (meters), occur in clusters, several meters tall, and are not stratabound in specific sandstone beds. Many set 4 fractures show evidence of reactivation in shear with both strike and dip senses of movement. For both types of reactivated set 4 fractures, shear reactivation is evidenced by wing cracks (**Figure 6b**). Offset bedding planes indicate bed-normal displacements ranging from 1-13 cm. A few set 4 fractures have subhorizontal striations indicating a strike-slip sense of shear. Linkage patterns by oblique segments exposed along pavements within Domain III are also compatible with strike-slip shear. Therefore, set 4 fractures seem to have formed by linkage and propagation of set 1 fractures, and were

subsequently reactivated in both strike-slip and dip-slip motion.

Faults

Although bedding-perpendicular, opening-mode fractures are the most common fracture type in these outcrops, faults with striated surfaces are also present in all three domains (**Figure 6c, d**). Low-angle thrust faults are abundant, their orientation varying with structural position (**Figure 6d-f**). A few faults that shorten bedding cut bedding at a steeper angle, and such high-angle reverse faults are best seen in Domain III along the southern bank of the river. Fault kinematics of thrust and reverse faults are compatible with SW-NE shortening, perpendicular to the fold axis.. Bedding-parallel slip surfaces are abundant on the hillside outcrops and most likely indicate flexural slip; such faults could form as a results of folding.

Cross-cutting relationships and relative timing

Consistent cross-cutting relationships between different fracture sets indicate that set 2 fractures cut set 1, and that set 3 cuts both set 1 and set 2 (**Figure 7**). Thus, set 1 fractures are the oldest and set 3 are the youngest of these three opening-mode fracture sets. Set 4 fractures are inferred to be younger than set 1 based on the differences noted above and because their location within fracture clusters and prevalence of slip surfaces suggest that they formed by reactivation and linkage of set 1 fractures. Fault kinematics of bedding-parallel faults, thrust and reverse faults, and slip along set 4 fractures are compatible with the expected kinematics of folding (**Figure 6d**). Set 4 fractures cut or abut against low-angle thrust faults, and low-angle faults do not displace the blocks cut by set 4 fractures, indicating that set 4 postdates low-angle thrust faults (**Figure 6b**). Propagation and linkage that led to the formation of set 4 fractures is thus inferred to have occurred during folding, and reactivation in shear along set 4 fractures is the

youngest event. Due to a lack of observed cross-cutting relations, the relative timing of set 5 is unknown.

Fracture cements

Quartz is the most common and most abundant fracture cement in all fracture sets, although some of the fractures also have minor calcite cement (less than 10%). Most of the microfractures and narrow macrofractures are completely filled with quartz cement, yet some of the wider fractures preserve traces of primary porosity up to ~7% of the fracture volume (**Figures 7a, c**). Narrow fractures exhibit crack-seal texture (**Figure 7b, d**) that indicates quartz cement precipitation concurrent or synkinematic with fracture opening (Ramsay, 1980; Laubach, 1988). In wider fractures ($> \sim 0.15$ mm), crack-seal texture is restricted to fracture margins (**Figure 7a, c**).

Fluid inclusion analyses of fracture quartz from one set 3 fracture within the inferred upper portion of the Kakwa Member in Domain III reveals primary two-phase inclusions containing a hydrocarbon liquid and a gas phase. Homogenization temperatures of these inclusions range from 69°C to 124°C. These temperatures are minimum estimates of trapping temperatures and of fracture cement precipitation, confirming the deep-burial conditions of quartz precipitated within set 3. These temperature estimates are in accordance with temperatures in excess of 80 °C required for fast accumulation of quartz cement (Walderhaug, 1994, 1996). The presence of synkinematic quartz cement in fractures in the study area indicates these fractures were opened and cemented deep in the subsurface and not at near-surface conditions.

Fracture abundance

Along a one dimensional line of observation (scanline), fracture intensity is defined here

as the number of fractures per unit length. In cumulative-frequency log-log plots, cumulative frequency of kinematic aperture indicates aperture-size rank (i.e. 1 is the widest fracture, 2 is the second widest, etc.) divided by scanline length, and kinematic aperture indicates the width of the fracture (Marrett, 1999; Ortega et al., 2006) (Figure 7). Cumulative frequency has units of number of fractures per unit length, thus its inverse is the average spacing of the fractures. Because the data are normalized (scanline length is taken into account) for different locations, scanline lengths and fracture apertures are directly comparable (Ortega et al., 2006).

Fracture scaling analyses were performed for two macroscanlines from Domains I and II, and 38 microscanlines from eleven field samples across the fold (**Figure 8, Table 2**). In general, microfracture data presented in **Figure 8** show power-law distributions with rollovers at the smallest apertures (truncation), and the largest scales of observation (censoring) (see Hooker et al., 2015 for discussion of truncation and censoring in fracture aperture distributions). A general scarcity of data representing fractures with apertures between 0.01-0.1 mm in these scanline data may reflect an observation gap between data collected in outcrop (macrofractures) and through SEM imaging (microfractures). In most data sets in this study, the narrowest microfractures can be fitted by a power-law distribution that has a steeper slope than that of the medium-sized fractures (**Figure 8a**). The slope of the narrowest microfractures ranges between -0.5 and -4.2 (most are between -0.9 and -2.8), whereas that of the medium-sized fractures ranges between -0.2 and -0.8 (most are between -0.2 and -0.5). Intensities (y axis) of the narrowest microfractures show a range of up to 1 order of magnitude for the same fracture set within the same domain. In contrast, the intensity of medium-sized fractures is similar and well constrained in each case. For example, the intensity of 0.1 mm aperture fractures is ~0.02-0.04 fractures/mm, irrespective of the fracture set or structural position (**Figure 8a**). The two field samples collected along scanline

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4 357 1 (RD-1 and RD-5) show similar intensities of the narrowest microfractures for both set 1 and set
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10 359 Because of the outcrop orientation, the two macroscanlines measured in Domains I and II
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12 360 contain predominantly set 1 fractures. Extrapolation of power-laws fitted through the narrowest
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14 361 set 1 microfracture data in these two domains is not a good predictor of the intensity of
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16 362 macrofractures, probably because the steep slopes of the narrowest microfractures are not
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18 363 maintained for fractures wider than about 0.01 mm (**Figure 8**). Based on a large dataset, Hooker
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20 et al. (2015) found that, in many cases, fitting a slope of -0.8 to measured microfracture data can
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22 364 predict the abundance of macrofractures to a first approximation. Both field samples collected
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24 365 along scanline 1 show overlapping intensities for the narrowest microfractures if the data
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26 366 affected by truncation bias are excluded. If we extrapolate the abundance of macrofractures from
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28 367 the narrowest microfractures using a slope of -0.8, we obtain a reasonably good prediction for
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30 368 macrofracture abundance in this part of the fold (**Figure 8a**). For scanline 2, this method
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32 369 provides estimates of fracture intensity good to within half an order of magnitude.
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39 371 Effect of lithology on fracture intensity 40 41

42 372 To evaluate the effect of lithology on fracturing in the Cardium Formation, we examined
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44 373 the fracture intensities of three adjacent samples from three different lithological units within the
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46 374 Kawka Member of Domain III: 1) a bioturbated layer –sample CO-14-3, 2) a planar-laminated
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48 375 layer –sample CO-14-2, and 3) a hummocky cross-stratified layer –sample CO-14-1, as well as
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50 376 samples from the planar-laminated upper (CO-15-2) and bioturbated lower (CO-15-1) portions
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52 377 within the Kakwa Member of Domain I (**Figures 3 and 8b**). These samples were cut in two
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54 378 orthogonal orientations to capture the three main fracture sets. Fracture intensities of the same
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56 379 fracture set differ to within half an order of magnitude in different lithologies. For example, the
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intensity of set 1 fractures in Domain III is higher in the planar-laminated layer than in the bioturbated and hummocky cross-stratified layers. This same distribution is found in the lower and upper portions of the Kakwa Member within Domain I.

Most of our measurements were made in layers of similar thickness. The exceptions were the lower and upper portions of the Kakwa in Domain I, where the thickness of the bioturbated lower Kakwa package (considered one layer) is similar to that of the planar-laminated upper Kakwa (tens of layers). Our data show no correlation between fracture abundance and layer thickness because layers of similar thicknesses (i.e. Domain III lithologies; **Figure 8b**) have different fracture intensities. Thus, we infer that bioturbation is more important than layer thickness on governing fracture abundance in the Cardium Formation. The correlation we found of fracture intensity with rock type holds for fractures in different orientations, but fracture abundance may be higher or lower depending on which orientation is considered. Lithological differences in clay content and distribution could be responsible for the observed within-sample (cm- to m-scale) variability in abundance of the narrowest microfractures.

Fracture strain

Strain Analysis Methods

We measured the strain accommodated by opening-mode fractures in the direction parallel to the cross section (roughly perpendicular to the fold axis) using two methods. In method 1, fracture strain ϵ is defined as a measure of the extension along a one dimensional line of observation (the scanline):

$$\epsilon = \frac{L - L_0}{L_0} = \frac{\Sigma \textit{Aperture}}{\Sigma \textit{Spacing}}$$

where L is the final length and L_0 is the initial length. Using this method, fracture strain equals the sum of all apertures divided by the sum of all spacings. To avoid counting fractures that were contained in both micro- and macroscanline datasets, fracture populations in microscanlines were truncated at apertures >0.1 mm and in macroscanlines at apertures ≤ 0.1 mm. Fracture strain for each location was calculated as the sum of strain recorded by fracture sets 1-3 observed in the field.

In method 2, strain is calculated as an aggregate strain estimate based on fracture frequency versus aperture plots (Figure 7) following Marrett et al. (1992) and Marrett (1996):

$$\varepsilon = \zeta \left(\frac{1}{b} \right) f_{\max} D_{\max}$$

where ζ is the Rieman zeta function, b is the slope of the power-law distribution function, $D_{\max}$ is the maximum fracture aperture, and $f_{\max}$ is the fracture frequency of $D_{\max}$. Following Marrett (1996), ζ is approximated by:

$$\zeta(x) \cong 1 + 2^{-x} + 3^{-x} + \frac{x+7}{2(x-1)} 4^{-x}$$

For strain calculations using method 2 we assumed a constant slope of -0.8, and used power-law distributions fitted through the narrowest microfractures. Fitting a slope of -0.8 through microfractures results in good predictions of macrofractures in our dataset, and can help overcome strain biases derived from undersampling of macrofractures in scanlines of limited length.

Strain Results

Strain calculated via method 1 using microfractures <0.1 mm in aperture ranges between 1.17×10^{-3} and 9.39×10^{-3} , whereas strain calculated using macrofractures >0.1 mm in aperture is

~6.7x10⁻³ for both macroscanlines. Strain calculated using method 2 ranges between 5.3x10⁻³ and 3.2x10⁻² when D_{max} is 0.1 mm, and 8.4x10⁻³ and 5.2x10⁻² when D_{max} is 1 mm (**Table 3**). Fracture strain shows no consistent variation with structural position in the fold, i.e. although strain magnitudes vary depending on the calculation method, relative strain values are similar across the fold regardless of structural position (**Figure 9; Table 3**). For example, fracture strain in the planar-laminated Kakwa Member is higher in the steeper dipping Domain III than in Domain I, but fracture strain for the Karr Member is higher in Domain I compared to Domain III. The large range in fracture strain within the Kakwa Member of Domain II suggests that local variations in fracture strain are larger than variations among different domains of the fold. Variations between different lithologic units of the same member in the same domain are equally large.

KINEMATIC MODEL AND STRAIN DISTRIBUTION

In the Red Deer River area, the spatial and temporal relationships between fractures and complex, seismic-scale structures (folds, faults) remain ambiguous. Geometrically valid framework models that adhere to geologic principles can be used to explain both the relationship of deformational components and their evolution throughout the deformation. For example, opening-mode fractures may accommodate bed-parallel, outer-arch extension associated with folding, resulting in a correlation between folding strain and fracture strain (Beekman et al., 2000; Tavani et al., 2014). In our study, we constructed kinematic models to 1) validate the structural interpretation of regional structures, 2) explore possible deformation styles that could have produced present-day structures, and 3) calculate bed-parallel extension in the tectonic transport direction during development of the fold-and-thrust belt. These strain calculations serve as a basis for predicting fracture strain associated with this deformation. Predicted strain distribution and timing relative to fold-and-thrust development were then tested by comparing

results to the distribution of fracture strain and timing obtained from field and petrographic observations.

Model construction

The basis of our numerical kinematic model were two cross-sections through the anticlinal structure exposed in the Red Deer River area, located 3 km apart (**Figures 1b and 10**). These cross-sections were based on cross-sections by Kubli and Langenberg (2002), with additional data by Ollerenshaw (1966). The Kubli and Langenberg (2002) cross-section incorporates surface geologic, stratigraphic, seismic, and subsurface data from 16 wells. Our cross-sections use surface geologic and structural data and stratigraphic information from Ollerenshaw (1996) for structures shallower than approximately 1500 meters depth, and Kubli and Langenberg (2002) cross-section for structures at depths between approximately 1500 and 5000 m. Our cross-sections were retrodeformed, line-length balanced, then forward modeled using Midland Valley Move 2014, a 2D structural restoration software. The cross-sections were verified to be line-length balanced and kinematically admissible in an iterative effort through reconstruction and line adjustment. The final version of the reconstructed sections are balanced and suited for restoration using Move's simple shear algorithm in 3D (**Figures 10a-b**).

Ollerenshaw's (1966) geologic map was draped on a digital elevation model (DEM), and both cross-sections were placed in a 3D space (**Figure 1b**). 3D surfaces were drawn across the two retrodeformed cross-sections by connecting equivalent horizon and fault lines using the "create surface from lines" algorithm of Move (**Figure 11**). The new cross-sections have five major faults numbered in the sequence of restoration, i.e. in reverse chronological order (**Figure 10**). Note that although cross-section B-B' contains a sixth fault (fault 4'), this fault is not present in cross section A-A', therefore it could not be extrapolated. Our initial 3D

retrodeformed model is thus a pseudo-3D model. The evolution of the Red Deer River anticline was subsequently modeled from the original starting point with horizontal bedding to the present-day structural geometry through progressive stages of faulting that proceeded from the hinterland to the foreland (**Figures 12 and 13**).

Due to limitations in available algorithms in Move 3D, we were not able to implement flexural slip as a deformation mechanism. Moreover, Move does not take into account layer thickness for strain calculations. In order to avoid differences in strain that would result from higher outer-arc than inner-arc extension we chose layers of similar thickness that are at comparable stratigraphic levels on both limbs of the anticline for strain analysis. During forward modeling, strain tensors were calculated for the top of the Cardium Formation using Move's strain tool. The strain calculated by this tool results from displacements imposed on the layer of interest by the thrust-and-fold development in the underlying units. No strains are imposed on the lateral boundaries of the model. Currently, Move does not isolate the elongation parallel to bedding and thus orthogonal to fractures oriented perpendicular to bedding but it calculates strain in terms of principal strain directions. Because bedding generally dips less than 45°, fracture-opening strain was measured using the horizontal strain component e_{xx} , and strain was calculated for both incremental steps and cumulative cases (**Figures 12 and 13**).

Strain patterns and magnitudes from models

Distribution of Incremental Maximum Horizontal Elongation e_{xx}

Magnitudes of the maximum horizontal elongation e_{xx} were calculated for multiple stages of displacement along each fault relative to the previous stage of deformation, i.e. resetting the strain to zero after each deformation stage. Strain from earlier stages is thus not included in the strain of the following deformational stage in **Figure 12**. Displacement along fault 5 is relatively

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4 489 small and does not affect the strain distribution around the eventual location of the Red Deer
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7 490 River anticline (**Figure 12a**). Fault 4 is a blind fault that does not cut through Cardium strata.
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9 491 Instead, strata above this fault are passively moved, and this displacement causes the highest e_{xx}
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12 492 strain to be concentrated around the future location of the Red Deer River anticline at this stage
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14 493 (**Figure 12b**).

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17 494 Displacement along fault 3 (termed Burnt Timber thrust fault in Ollerenshaw's (1966)
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19 495 cross section) is relatively large and has been subdivided into several steps. This fault is a splay
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22 496 off a larger structure at depth, the Blue Hills thrust, which displaces the Paleozoic section with a
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24 497 displacement on the order of ~25 km (Peter Fermor, personal communication). During the early
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27 498 stage of deformation along the Burnt Timber thrust, high elongation areas are located in the
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29 499 limbs of the evolving Red Deer River anticline (**Figure 12b**), but as displacement continued, the
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32 500 locations of high and low strain concentrations changed. High strain (indicated by red)
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34 501 subsequently decreased (blue colors) as the layer passed through the anticline and flattened into
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36 502 the forelimb (**Figures 12c-e**). Elongation is concentrated in the forelimb of the Red Deer River
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39 503 anticline in stage c, and in the backlimb during stage d. As deformation progresses, the highest
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41 504 elongation is consistently concentrated near the deformation front, whereas strain around the
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44 505 evolving Red Deer River anticline is highly variable throughout the deformation. Displacement
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46 506 along faults 2 and 1 resulted in elongation east of the Red Deer River anticline in the layers
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49 507 above the James River anticline, which is now largely eroded, while the Red Deer River anticline
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51 508 itself remained relatively undeformed (**Figures 12f-h**). These models predict that most of the
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54 509 horizontal elongation experienced by the Cardium Formation in the Red Deer River anticline
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56 510 occurred in the early stages of displacement along fault 3, with a pronounced phase of extension
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58 511 in the backlimb of the Red Deer River anticline during stage d (**Figure 12d**).

Distribution of Cumulative Horizontal Elongation e_{xx}

Cumulative maximum horizontal elongation was calculated for stages a through h in **Figure 13** by adding together all strain increments relative to the initial undeformed state. Thus, each stage represents the cumulative strain the rock has experienced. Similar to the non-cumulative strain case, displacement along fault 3 caused the highest elongation to initially occur in both limbs of the Red Deer River anticline (**Figure 13b**), and further displacement along this fault resulted in elongation of the backlimb of the Red Deer River anticline (**Figure 13c**). As displacement continued along fault 3, areas that were highly elongated developed into lower-strain areas, indicating that both originally-elongated strata underwent shortening as deformation progressed and that folded layers flattened out as they migrated through the fold hinge (**Figure 13d-e**). The relatively small displacement along fault 2 did not have any noticeable effect on the strain distribution of Cardium strata (Figure 13f), whereas slip along fault 1 caused elongation of the Cardium strata to the east of the Burnt Timber thrust above the James River anticline (**Figure 13g-h**).

Therefore, the forelimb of the Red Deer River anticline undergoes a distinct phase of cumulative extension during stage c. From stage d onward, somewhat higher elongation is observed in the backlimb (green colors) compared to the forelimb (blue colors) of the Red Deer River anticline. This difference is less than 0.1. Within the forelimb, somewhat lower strains are observed toward the crest, and higher strains toward the trough of the Red Deer River syncline which is located east of the anticline. The final cumulative strain distribution shows slightly larger elongation in the backlimb of the Red Deer River anticline than in the forelimb, but cumulative strain magnitudes are similar on both limbs. The areas of highest elongation are located in the now eroded Cardium Formation east of the Red Deer River anticline.

DISCUSSION

Dechesne and Villeaux (2000) identified three main sets of fractures in the Red Deer River anticline outcrops, cross-fold (perpendicular to the fold axis), strike-parallel (parallel to the fold axis), and cross-strike (oblique to the fold axis), their orientation varying across the fold with structural position. They concluded that local stresses were the main control on fracture orientation in the area, although fracture orientations seem to have been overprinted by the regional fracture sets present in the Plains to the east of the fold belt. These authors did not conduct a fold-test, nor did they provide timing estimates of bed-perpendicular fractures with respect to the formation of the fold.

In this study, we identified four main bedding-perpendicular sets, two of which share the same orientation. Fractures are not parallel and orthogonal to the fold axis, but are slightly oblique (sub-parallel and sub-perpendicular). Cross-cutting relationships indicate at least some fractures are fold-related but some might predate folding. Fault kinematics of bedding-parallel faults, thrust and reverse faults, and slip along set 4 fractures are compatible with the expected kinematics of folding (**Figure 6d**). Set 4 fractures cut or abut against low-angle thrust faults, and low-angle faults do not displace the blocks cut by set 4 fractures, indicating that set 4 postdates low-angle thrust faults (**Figure 6b**). Propagation and linkage that led to the formation of set 4 fractures is thus inferred to have occurred during folding, and reactivation in shear along set 4 fractures is the youngest event. When bedding is rotated to horizontal, and fractures are rotated accordingly (fold test), the fracture orientations cluster more tightly than the unrotated (measured) fracture orientations, suggesting that sets 1-3 formed when bedding was flat, prior to folding of the Red Deer River anticline, and were rotated during folding. However, it is conceivable that fracture orientation can be influenced by the layer anisotropy, with fractures

forming perpendicular to bedding, regardless of bedding dip. Thus, bedding-perpendicular fractures may have formed either prior to or during folding.

The Canadian Rocky Mountains have undergone a protracted history of SW-NE contraction (Bally et al., 1966; Eisbacher et al., 1974; Price, 1994), which continues to the present day. Opening-mode fractures analyzed in this study thus record a relatively small amount of extension (~1-10%) superimposed on overall regional contraction that resulted in km-scale reverse faults. Because of this prolonged stress field, the main orientation of mode I fractures in the Canadian Rocky Mountains and the West Central Alberta Basin is predicted to be SW-NE, which is confirmed by fractures in outcrop and core, and borehole break-out data (Bell and Gouh, 1979; Dechesne and Veilleux, 2000). In our study area, after unfolding the strata, fracture set 1, the oldest set, and set 4, the youngest set, are oriented subparallel to the present-day maximum horizontal in-situ stress trend of 45-60° (**Figure 5b**). In some wells east of the fold-and-thrust belt an additional fracture set striking 330° has been observed (Dechesne and Veilleux, 2000). Fracture set 3 in our study area strikes ~300° differing by ~30° from this regional set. No regional set of fractures has been reported in the orientation of set 2 (~0-30°).

The homogeneity we observe in macro- and microfracture intensity across the Red Deer River anticline, especially for fractures >0.1 mm in aperture, is compatible with either a pre- or synfolding origin of fracture sets 1-3 (**Figure 8**). We infer that bioturbation is more important than layer thickness on governing fracture abundance in the Cardium Formation. Bioturbation generally marks slightly deeper water sandstone deposits that have higher proportions of compliant and fine-grained materials. These slight compositional changes along with the textural homogenization accompanying bioturbation result in different mechanical properties from clean

and bioturbated sandstones and a widely recognized propensity for such sandstones to have lower fractures abundance (e.g., Nelson, 1985; Laubach et al., 2009).

Pre-folding fractures can become reactivated during folding leading to re-opening, shearing, and linkage of existing fractures, and formation of new macro- and microfractures (Dunne 1986; Laubach, 1988; Davatzes and Aydin, 2003; Bergbauer and Pollard, 2004). Such a reactivation model was proposed for similar folds and rock types by Inigo et al. (2012). The progressive addition of fractures that would result from this model is compatible for long-lived partly open large fractures (e.g., Laubach and Ward, 2006). If such fractures were cemented concurrently or synkinematically with fracture opening they might record past fold-related strain, even if they are in parts of the fold not currently in an area of high curvature (for example, former hinges that have been unfolded). Fractures in the Cardium Formation are generally cemented with synkinematic quartz cement (**Figure 7**), which would prevent closure of fractures when stress and strain conditions change in the fold during progressive evolution of the thrust sequence.

Two kinematic models have previously been proposed to predict the occurrence of opening-mode fractures in folds: fixed-hinge (De Sitter, 1956) and active-hinge (Suppe, 1985) models. Fixed-hinge folding implies no material migration across axial surfaces while limbs rotate about a fixed axial surface (De Sitter, 1956; McConnell, 1994; Salvini and Storti, 2004; Gosh and Saha, 2005; Mercier, 2007), while active-hinge folding implies material migration across axial surfaces relative to a stable position of a ramp (Suppe et al., 1992). Both mechanisms can produce folds having a similar appearance, but the deformation patterns would be different (Fisher et al., 1992; Salvini and Storti, 2001). For a fold where the hinge is stationary, strain is contractional in the inner arc, and extensional in the outer arc of the folded layer. Therefore, the

highest extensional strain is expected to be accommodated by opening-mode fractures in the hinge of a fold. Thus, fixed-hinge models can be used to predict spatial distribution of deformation as a function of layer curvature (Murray, 1968; Decker et al., 1989; Schultz-Ela and Yeh, 1992; Lemiszki et al., 1994; Lisle, 1994; Hennings et al., 2000; Fisher and Wilkerson, 2000). In contrast, active-hinge folding results in hinge migration where the distribution of high deformation intensity and associated strain tracks the hinge migration path (Suppe, 1983; Fischer et al., 1992; Salvini and Storti, 2004).

Strain calculations using Move software for the Red Deer River anticline predict that fault-related fold development results in a complex strain path throughout deformation consistent with active-hinge folding (**Figures 12 and 13**), and with extensional strain localized primarily in the fold limbs. Strain calculation provides information about the timing of strain that could lead to fracture opening relative to the thrust evolution. While incremental strain provides a proxy for predicting fracture opening at specific stages of fold and thrust evolution, the *cumulative* strain can be considered a proxy for strain accommodated by fractures cemented by mineral cement concurrently or synkinematically with fracture opening, thus preserving fracture aperture.

Fracture strain calculated from scanlines and from kinematic models are comparable in magnitude. Incremental e_{xx} values modeled with Move are as high as 0.1 around the Red Deer River anticline (**Figure 12**) but in most cases strain values are within the same order of magnitude as those derived from microfractures in the field. Strain magnitudes derived from microfracture and macrofracture data from field samples are between 0.001 and 0.009 using method 1, and 0.005 and 0.03 using method 2 when D_{max} is 0.1 mm. In contrast, strain calculated based on fractures up to 1 mm in aperture using method 2 is generally one order of magnitude greater, between 0.008 and 0.05. This difference highlights the scale-dependency of strain

measurements (Marrett, 1996). We recognize that the integrated strain calculated based on a $D_{\max}$ = 1 mm probably overestimates true fracture strain as indicated by an overall scarcity of medium-sized fractures in these samples. In comparison to the strain results derived from Move, the fracture strain observed in the planar-laminated upper portion of the Kakwa Member shows little variation across the three domains (Figure 8), with Domain II, the upper section of the forelimb, showing somewhat lower strain than the lower forelimb (Domain III) and the backlimb (Domain I). Although the number of strain measurements in the upper portion of the Kakwa Member is not evenly distributed across the three domains (with only one measurement in Domain III), we find that the strain distribution is consistent with the cumulative strain calculations in Move (**Figure 9**). Similar extension magnitudes, on the order of ~1-10%, have been reported for many geological examples (e.g., Narahara and Wiltchko, 1986; Jamison, 1997; Gross and Engelder, 1995; Inigo et al., 2012), irrespective of the tectonic setting. The similarity between fracture strain calculated using methods 1 and 2 for equivalent scales highlights the robustness of the method.

Kinematic models indicate that the largest incremental extension would occur in the backlimb of the Red Deer River anticline early in the development of the Burnt Timber thrust (**Figure 12, stage d**) when the forelimb undergoes a shortening phase. If fracturing were associated with folding, most opening-mode fractures would thus form during this stage. However, cumulative strains (**Figure 13**) in both limbs of the anticline are roughly similar from stage d onward, with somewhat higher extension in the backlimb and lower extension in the forelimb close to the crest of the fold. These models thus predict that strain recorded by fold related, bed-perpendicular opening-mode fractures would be similar in both limbs of the fold. Cumulative strain calculated by Move is thus compatible with our strain measurements using scanlines and indicates opening and cementation of fractures during early folding. Such a rather uniform strain across the fold could also suggest fracture formation prior to folding and subsequent

reactivation. In Alberta, we found evidence for faulting and reactivation in shear of probable earlier-formed opening-mode fractures. Reactivation of fractures is predicted by geomechanical models and extensive evidence for reactivation is found in outcrop (e.g., Guiton et al., 2003; Smart et al., 2009). In folds of about the same age in Wyoming, Bergbauer and Pollard (2004), found evidence that the orientation of pre-existing fractures controlled the internal structures and the orientation of fractures that form later in the deformation history. Tall, thoroughgoing fractures may be susceptible to reactivation (Becker and Gross, 1996). The linkage geometry we observed of through-going set 4 fractures together with evidence of shear offset suggests that they formed by reactivation and linkage of earlier set 1, bedding-confined fractures. We speculate that set 1 fractures contributed to the mechanical weakness that may have influenced subsequent fracturing during multiple phases of brittle deformation.

CONCLUSIONS

Opening-mode fractures are prominent in surface exposures of the Cardium Formation within the Red Deer River anticline. Three main opening-mode fracture sets were identified, and fracture spacing and aperture data were collected along both macro- and microscanlines. These data indicate that variations in fracture intensity between the fracture sets are small among the backlimb (Domain I), upper forelimb (Domain II), and lower forelimb (Domain III) across the fold, yet the intensity is slightly lower in the upper forelimb versus the backlimb and lower forelimb of the anticline, which are both higher (and equal) in intensity.

The large-scale fold geometry of the Red Deer River anticline can be explained by kinematic restoration of regional-scale cross-sections and forward modeling. These kinematic models suggest that this anticline developed by active-hinge migration resulting from slip along the underlying Burnt Timber thrust. A comparison of measured fracture strain with the kinematic

model predictions suggests that the observed fracture strain distribution across the three structural domains is consistent with fracture formation during early development of the Red Deer River anticline. However, a pre-folding origin of the fractures cannot be excluded because the differences in measured fracture strain across the three structural domains of the fold are small, and layer-to-layer variation in folding strain within each domain are large.

Formation of fractures prior to, or at the onset of, folding is consistent with systematic angular relations between the three bed-bounded, high-angle-to-bedding sets of opening-mode fractures, which suggests that the fractures were progressively rotated as bedding planes were folded. Commonly observed through-going fractures formed by propagation and linkage are associated with wing cracks and show both strike-slip and dip-slip senses of shearing, which are kinematically compatible with folding. This suggests that through-going fractures were reactivated in shear during the later stages of fold development.

Because a different strain distribution is predicted for different parts of the fold at different times during deformation, the numerical simulations performed in this study demonstrate that fold evolution within deforming fold-and-thrust-belts can result in complex distributions of bedding-parallel extension, and thus in the formation of opening-mode fractures, in both space and time.

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FIGURE CAPTIONS

Figure 1: a) Tectonic subdivisions in the southern Canadian Cordillera, modified from Hardebol et al. (2009). **b)** Geologic map of Ollerenshaw (1966) showing the location of cross-sections A-A' and B-B'. WCSB: Western Canada Sedimentary Basin.

Figure 2: Stratigraphic column of the Alberta Foothills and Front Ranges (Gordy et al., 1975).

Figure 3: a) Panoramic view of the Red Deer River anticline (N 51°39'30.4''W 115°17'12.2'').

The anticline is subdivided into three structural Domains I-III. Sandstone members of the Cardium Formation in the study area are traced in white. Yellow dots (Sc-1 and Sc-2) indicate the location of the two macroscanlines measured in the field. Red dots and labels indicate the location of the hand samples collected for scaling analyses. **b)** Macroscanline Sc-2. **c)** Macroscanline Sc-1. **d)** Close-up of Macroscanline Sc-2. Yellow line indicates the scanline. **e)** Close up of sample CO-14-2 showing vertical fractures. **f)** Location of sample RD-2 within Domain III on the south side of the Red Deer River. **g)** Location of samples CO-14-1, CO-14-2, and CO-14-3 stratigraphically above RD-2 within the same outcrop.

Figure 4: Schematic block diagram showing the orientation of the different fracture sets across the fold. Not to scale.

Figure 5a: Poles to fracture planes of the most prominent opening-mode fracture sets from different parts of the fold. Great circles indicate the orientation of bedding. Data shown in gray were measured in macroscanlines. Lower hemisphere projection.

Figure 5b: Poles to fracture planes of the most prominent opening-mode fracture sets in different parts of the fold after bedding is rotated to horizontal. Data shown in gray were measured in macroscanlines. Lower hemisphere projection.

Figure 6: a) Through-going fractures (set 4) within the Kakwa Member in Domain II. Set 4 fractures probably formed by linkage and propagation of set 1 fractures (note the fracture clusters along which through-going fractures are concentrated). **b)** Inset showing wing cracks and offset layers that indicate dip-slip movement of ~ 1cm. **c)** Set 4 fracture cutting across a low-angle thrust fault. **d)** Fault kinematics of representative low-angle thrust faults in Domain II. **e)** Right-lateral, strike-slip reactivation of set 4 fracture in Domain II. **f)** Representative low-angle thrust, and reactivation in shear and dip-slip, of steeply dipping set 4 fractures in Domain III south of the river. Fault kinematics are compatible with shearing related to folding. TF = thrust fault; RLSS = right-lateral strike-slip fault; LLSS = left-lateral strike-slip fault. Arrows indicate slip direction of the hanging wall.

Figure 7: a) SEM-CL image of set 3 fracture fill showing the euhedral nature of the quartz crystals and preserved porosity. Sample RD-2B-A. **b)** Set 3 fracture showing crack-seal texture along the fracture wall and euhedral cement fill away from the wall. Sample RD-2B-D. **c)** SEM-CL image of set 1 fracture fill showing the euhedral nature of the quartz crystals and preserved porosity. Sample CO-14-2a-d. **d)** SEM-CL image of example cross-cutting and abutting fractures. Note the crack-seal texture of the narrow set 1 (vertical) and set 3 (horizontal) fractures, whereas the wider set 1 fracture (upper left) is mostly infilled with euhedral quartz cement. Sample CO-14-2a-d. CS = crack-seal; E = euhedral quartz; P = porosity.

Figure 8a: Cumulative frequency versus kinematic aperture plots of microfracture sets 1-3 in the three structural domains. Open symbols are for macrofractures measured in scanlines 1 and 2. The red line has a slope of -0.8. Extrapolation of microfracture data using a constant slope of -0.8 gives a good prediction of macrofracture abundance. Dashed gray lines mark the intensity of 0.1 mm aperture fractures.

Figure 8b: Cumulative frequency versus kinematic aperture plots of microfractures showing the effect of lithology on fracture abundance.

Figure 9: Fracture strain plot according to structural position and lithology. Symbols outlined in black are for calculations using macroscanlines >0.1 mm in aperture.

Figure 10: Balanced and simplified cross-sections A-A' (a) and B-B' (b) used for the construction of the 3D kinematic model. Circled numbers indicate fault numbers referred to in the text. Colors in the legend indicate the top of each formation. The stratigraphic column has been simplified. c) Original cross-section of Ollerenshaw (1966), which corresponds to the trace of cross-section B-B'. Stratigraphic sequence and colors as in Figure 1b.

Figure 11: Retrodeformed cross-section B-B'. The now-eroded strata around the hinge of the Red Deer River anticline and above the Burnt Timber thrust are shown in dashed lines.

Figure 12: Incremental evolution of the horizontal elongation component e_{xx} (non-cumulative) for the Cardium Formation. Stratigraphic horizon colors are the same as in Figure 10. Color scale bars indicate magnitude of e_{xx} in each case for the deforming top of the Cardium Formation.

Figure 13: Evolution of the horizontal elongation component e_{xx} (cumulative) for the Cardium Formation. Stratigraphic horizon colors are the same as in Figure 10. Color scale bars indicate magnitude of e_{xx} in each case for the deforming top of the Cardium Formation.

Table 1: General attributes of fracture sets based on data collected in Domain II.

Table 2: Best-fit equations and R^2 values of the narrowest and medium-sized (apertures between 0.01-0.1 mm) microfractures.

Table 3: Strain values calculated using methods 1 and 2.

Figure 1

Figure 1

Figure 2

Figure 2

Figure 3

Figure 3

Figure 4

Figure 4

Figure 5a

Figure 5a

Figure 5b

Figure 5b

Figure 6

Figure 6

Figure 7

Figure 7

Figure 8a

Figure 8a

Figure 8b

Figure 8b

Figure 9

Figure 9

Figure 10

Figure 10

Figure 11

Figure 11

Figure 12

Figure 12

Figure 13

Figure 13

Table 1: General attributes of fracture sets based on data collected in Domain II.

Attribute/ Set	Set 1	Set 2	Set 3
Strike	NE-SW	N-S	WNW-ESE
Strike after unfolding	NE-SW	N-S	WNW-ESE
Dip	Bedding-perpendicular	Bedding-perpendicular	Bedding-perpendicular
Relative Age	Oldest	Younger than set 1, older than set 3	Younger than set 1 and set 2, older than set 4
Quartz cement	Yes	Yes	Yes
Associated microfractures	Yes	Yes	Yes

Table 2: Best-fit equations and R² values of the narrowest and medium-sized mic

Sample	Scanline length (mm)	Micro	
		narrowest microfractures	
co-15-1a-1	64.73	$y = 4E-07x^{-2.139}$	R ² = 0.9168
co-15-2a-1	60.37	$y = 1E-05x^{-1.635}$	R ² = 0.9371
co-15-2b-1	87.15	$y = 6E-05x^{-1.208}$	R ² = 0.969
co-15-1a-2	64.73	$y = 2E-09x^{-3.099}$	R ² = 0.9843
co-15-2a-2	60.37	$y = 3E-05x^{-1.534}$	R ² = 0.9638
Rd-4b-all	244.13	$y = 5E-12x^{-4.211}$	R ² = 0.9852
co-15-2b-2	87.15	$y = 0.0002x^{-1.093}$	R ² = 0.9656
co-15-2a-3	60.37	$y = 0.0015x^{-0.724}$	R ² = 0.9853
co-15-2b-3	87.15	$y = 1E-05x^{-1.737}$	R ² = 0.9672
Rd-4a-all	48.32	$y = 2E-05x^{-1.659}$	R ² = 0.9213
co-15-3b-1	66.47	$y = 1E-05x^{-1.533}$	R ² = 0.7489
co-15-3a-1	104.14	$y = 4E-10x^{-3.454}$	R ² = 0.8848
co-15-3a-2	104.14	$y = 1E-05x^{-1.779}$	R ² = 0.9656
co-15-3a-3	104.14	$y = 3E-09x^{-3.113}$	R ² = 0.9553
co-15-3b-3	66.47	$y = 0.0024x^{-0.698}$	R ² = 0.9511
Rd-1b-all	168.53	$y = 1E-05x^{-1.687}$	R ² = 0.9876
Rd-5b-all	136.01	$y = 9E-09x^{-2.817}$	R ² = 0.9865
Rd-1a-all	26.21	$y = 1E-07x^{-2.493}$	R ² = 0.9721
Rd-5a-all	40.77	$y = 2E-06x^{-2.12}$	R ² = 0.8903
co-14-1a-1	178.88	$y = 9E-07x^{-2.157}$	R ² = 0.9799
co-14-2a-1	133.06	$y = 4E-05x^{-1.54}$	R ² = 0.9871
co-14-2b-1	101.14	$y = 3E-07x^{-2.045}$	R ² = 0.9661
co-14-3a-1	258.35	$y = 7E-11x^{-3.816}$	R ² = 0.9832
co-16-1a-1	103.03	$y = 0.0003x^{-1.016}$	R ² = 0.9835
co-16-1b-1	60.80	$y = 0.0003x^{-0.944}$	R ² = 0.9381
Rd-2b-1	200.62	$y = 5E-06x^{-1.856}$	R ² = 0.9639
co-14-1a-2	178.88	$y = 0.0011x^{-0.505}$	R ² = 0.9878
co-14-2a-2	133.06	$y = 0.0005x^{-1.208}$	R ² = 0.9893
co-14-2b-2	101.14	$y = 8E-09x^{-2.686}$	R ² = 0.9525
co-16-1a-2	103.03	$y = 5E-09x^{-2.826}$	R ² = 0.9089
co-16-1b-2	60.80	$y = 5E-09x^{-2.826}$	R ² = 0.9089
Rd-2b-2	200.62	$y = 2E-05x^{-1.567}$	R ² = 0.9739
co-14-1a-3	178.88	$y = 0.0003x^{-0.862}$	R ² = 0.9299

Table 3:

Scanline	Strain, Method 1	Strain, Method 2	
		Dmax = 0.1 mm	Dmax = 1 mm
CO-14-1a	0.0026	0.0091	0.0144
CO-14-2a	0.0086	0.0329	0.0521
CO-14-3a	0.0029	0.0098	0.0155
CO-15-1a	0.0018	0.0058	0.0091
CO-15-2a	0.0067	0.0167	0.0265
CO-15-3a	0.0076	0.0211	0.0334
CO-16-1a	0.0023	0.0088	0.0139
RD-1A-Sc-all	0.0044	0.0173	0.0274
RD-1B-Sc-all	0.0026	0.0081	0.0128
RD-2A-Sc-all	0.0070	0.0098	0.0155
RD-2B-Sc-all	0.0094	0.0219	0.0347
RD-4A-Sc-all	0.0035	0.0115	0.0183
RD-4B-Sc-all	0.0012	0.0058	0.0091
RD-5A-Sc-all	0.0048	0.0150	0.0238
RD-5B-Sc-all	0.0017	0.0081	0.0128
EU-Sc-1 (macro)	0.0068	0.0081	0.0128
EU-Sc-2 (macro)	0.0060	0.0053	0.0084