Natural hydraulic fracturing of tight-gas sandstone reservoirs, Piceance Basin, Colorado

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ABSTRACT

Natural fractures form preferred pathways for basin fluid flow and associated heat and mass transport. In gas sandstone reservoirs with low matrix permeability, fractures provide flow paths between organic-rich source and reservoir layers during gas charge, and between matrix pores, hydraulic fractures, and the well bore during production. While the formation of natural fractures has previously been associated with gas generation and pore-fluid pressure increase through a process referred to as natural hydraulic fracturing, other driving mechanisms such as stress changes by tectonic or exhumation processes remained viable alternatives. To test whether these mechanisms contributed to fracture development, we investigated the spatial and temporal distribution of fracture formation and its relationship to gas generation, migration, and charge in sandstone of the Cretaceous Mesaverde Group across the entire production interval on a basinwide scale. Using fluid inclusion microthermometry of crack-seal fracture cement formed concurrently with fracture opening, we observed temperature trends that, when compared with temperature evolution models of the formation, date fracture formation between 41 and 6 Ma in the northern and between 39 and 6 Ma in the southern Piceance Basin. The onset of fracture formation 20–30 m.y. prior to maximum burial eliminates changes in stress state associated with exhumation as a mechanism for triggering the onset of fracture formation. Instead, calculated paleo-pore-fluid pressures of 40–90 MPa (5800–13,000 psi) during fracture opening and the presence of methane-rich inclusions in fracture cement suggest that fracture formation was aided by high pore-fluid pressures during gas generation in organic-rich shales and coals and associated charging of adjacent and interlayered sandstone reservoirs. A 10–20 m.y. age progression in the onset of fracture formation from deeper to shallower horizons of the Mesaverde Group is consistent with gas generation and onset of fracture formation activated by burial temperature with limited upward migration of gas at this stage of reservoir evolution. This age progression with depth is inconsistent with fracture formation triggered by changes in stress conditions associated with tectonic or structural processes expected to affect the entire formation synchronously. Our observations are thus most consistent with fracture formation by natural hydraulic fracturing in response to gas generation in interbedded source layers and reservoir charge. Based on widespread observations of fractures with similar structural and diagenetic attributes, we consider natural hydraulic fracture formation in response to thermocatalytic gas generation to be a fundamental mode of brittle failure in otherwise structurally quiescent basins.

INTRODUCTION

Fractures and faults control transport of fluid and heat in the upper crust (Bredehoef and Norton, 1990; Eichhubl and Boles, 2000b). Processes leading to the formation of fractures and fracture networks are thus essential for understanding the thermal and chemical evolution of the upper crust and its mechanical properties. Fractures are also relevant for the development of water and fossil energy resources, and the secure geologic storage of CO2 and radionuclides (National Research Council, 1996; Chiarantome et al., 2011). In unconventional tight-gas sandstone reservoirs, natural fractures play an essential role during well completion and production by connecting matrix pores to the wellbore through hydraulic and mechanical interaction with induced hydraulic fractures (Dahi-Taleghani and Olson, 2011). Natural opening-mode or sheared opening-mode fractures also play an essential role during gas charge of the reservoir, providing pathways for gas migration from source rocks to matrix pores (Laubach, 1988; Lorenz and Finley, 1991; Cumella and Scheevel, 2008; Olson et al., 2009). “Sweet spots” or zones of higher than average permeability have been attributed to the presence of open fractures (Surdam, 1997; Laubach, 2003; Cumella and Scheevel, 2008). Successful prediction of zones of increased fracture abundance provides an opportunity to minimize drilling and completion costs and the environmental footprint of production. Successful prediction of natural fracture occurrence and their hydraulic properties requires models of fracture formation that are based on realistic mechanical, hydraulic, and chemical principles and that can be tested against core, well log, and production data.

Since the seminal work by Hubbert and Willis (1957), Handin et al. (1963), and Secor (1965), it is now universally accepted that high pore pressure aids the formation of natural fractures under reservoir conditions, with opening-mode fractures forming when the pore-fluid pressure p exceeds the least principal compressive total stress σt, resulting in a tensile effective least principal stress σ′ (Engelder, 1985; Engelder and Lacazette, 1990; Lacazette and Engelder, 1992). Following Secor (1965) and including poroelastic components of the stress tensor into the total stress term, the condition for fracture formation can be expressed as

σ′ = (σ1 – p) < T0,

where T0 designates the tensile strength (where compressive stress and pore-fluid pressure are considered positive quantities). For a rock formation containing existing fractures, T0 will be a small quantity, simplifying the condition for fracture formation to

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The process of fracture formation by a natural increase in pore-fluid pressure has previously been referred to as natural hydraulic fracturing (Phillips, 1972; Engelder, 1985). In tight-gas sandstone reservoirs, it has been proposed that high pore-fluid pressures can be achieved during thermocatalytic gas generation in interbedded organic-rich shales and coals, and concomitant gas migration into the adjacent sandstone reservoir layers (Spencer, 1987; Lorenz and Finley, 1991; Cumella and Scheevel, 2008; Yurewicz et al., 2008). While thermocatalytic gas generation under deep burial conditions provides a mechanism for increasing pore-fluid pressure and thus generating conditions favorable for fracture formation (Tingay et al., 2013), we cannot rule out other mechanisms that would lead to a tensile effective least principal stress \( \sigma_3' \), and thus to fracture opening. Such processes include an increase in pore-fluid pressure by compaction disequilibrium and the reduction in confining stress and thus in the least principal total stress \( \sigma_3 \), associated with tectonic processes or exhumation (Currie and Nwachukwu, 1974; Narr and Currie, 1982; Engelder, 1985; Lorenz and Finley, 1991; Brown, 2005; English, 2012). Although not mutually exclusive, fracture formation driven primarily by a decrease in total stress \( \sigma_3 \), rather than by an increase in pore-fluid pressure, is not to be considered natural hydraulic fracturing.

Recent fluid inclusion studies on crack-seal fracture cements that formed concurrently with fracture opening have provided age constraints on fracture formation in tight-gas sandstone reservoirs. Based on a correlation of consistent temperature trends obtained using fluid inclusion microthermometry of crack-seal fracture cement with independently derived burial temperature history models of the reservoirs, Becker et al. (2010), for the East Texas Basin, and Fall et al. (2012), for the southern Piceance Basin, demonstrated that fracture formation was a prolonged process over time spans of 35 and 48 m.y., respectively. Following Becker et al. (2010) and Fall et al. (2012), we note that the onset of fracture opening in the East Texas and Piceance Basins coincided approximately with the onset of gas maturation under conditions approaching, but prior to, maximum burial, and below the depth range of significant mechanical or chemical compaction, thus eliminating compaction disequilibrium as the dominant mechanism increasing pore-fluid pressure at the time of fracture initiation. Instead, methane-rich fluid inclusions in fracture cement indicated fracture opening during gas generation under elevated pore-fluid pressures consistent with fracture opening driven by gas generation. The apparent onset of fracture formation prior to maximum burial also eliminated processes associated with exhumation and associated changes in confining stress (e.g., English, 2012) as the mechanism that initiated fracture formation. However, changes in stress state due to basin- or reservoir-scale tectonic or structural processes, such as a change in the tectonic stress regime at about the onset of Laramide regional deformation, as discussed by Lorenz and Finley (1991) for the Piceance Basin, could not be ruled out as a trigger for fracture formation because sampling was limited in stratigraphic and geographic extent, thus preventing a rigorous analysis of depth and basin-scale trends and processes.

This study extends the work of Fall et al. (2012) to test models of fracture formation and associated pore-pressure generation in the Cretaceous Mesaverde Group of the Piceance Basin by determining the timing of fracture opening relative to the burial evolution and pore-fluid pressure history for multiple reservoir depths on a basinwide scale. This expanded basinwide approach was designed to address the following questions: (1) When do fractures form relative to the onset of gas generation in potential source layers? (2) Does the timing of fracture formation reflect gas generation in deeper source layers and upward gas migration, resulting in fracture formation that is synchronous throughout the Mesaverde Group, or (3) is fracture formation in shallower reservoir layers delayed relative to deeper reservoir layers, reflecting later gas generation in shallower source layers with no or minimal upward gas migration? (4) Does the timing of fracture formation reflect spatial trends that follow reservoir-scale structural or basin-scale tectonic or exhumation processes? In addressing these questions, we attempt to test for a causal link between the processes of gas generation and migration and fracture formation.

To address these questions, we combined scanning electron microscopy–cathodoluminescence (SEM–CL) imaging of fracture cements with microthermometry and Raman microspectrometry of fluid inclusions trapped in the cements to constrain the fluid pressure and compositional evolution during fracture opening and cementation in the Piceance Creek and Love Ranch fields in the northern Piceance Basin, Colorado. Trapping temperatures of fluid inclusions observed in fracture cements were correlated with independently derived burial history models to determine the relative timing of fracture opening. Fluid pressure conditions during fracture opening were determined based on gas concentration measurements in fluid inclusions and equation-of-state calculations. This approach follows Becker et al. (2010) and Fall et al. (2012) and builds on earlier diagenetic fracture cement fluid inclusion studies (Vrolijk et al., 1988; Goldstein and Reynolds, 1994; Evans and Battles, 1999; Eichhubl and Boles, 2000a; Parris et al., 2003; Hanks et al., 2006; Laubach and Díaz-Tushman, 2009; Duncan et al., 2012), and crack-seal cement textural analysis (Laubach et al., 2004a, 2004b). The results from the northern Piceance Basin are compared to reanalyzed results from the southern Piceance Basin previously presented by Fall et al. (2012) to obtain a basinwide model of fracture opening in relation to gas generation and charge.

**GEOLOGIC SETTING**

The Piceance Basin is an intracratonic foreland basin in northwestern Colorado (Fig. 1A) that formed to the east of the Sevier orogenic belt during the Late Cretaceous to Eocene Laramide orogeny (Johnson and Nuccio, 1986; Johnson and Rice, 1990; Patterson et al., 2003). Except for a Late Cretaceous period of uplift, rapid subsidence during Cretaceous and Paleogene times resulted in a thick Upper Cretaceous to Paleogene sedimentary sequence. After maximum burial during late Oligocene–early Miocene times, uplift along the basin flanks associated with late Laramide top-to-the-west high-angle reverse faulting along the White River uplift (Lorenz and Finley, 1991) resulted in the modern asymmetric basin shape with gently dipping flanks on the west and a steep flank on the east side (Fig. 1B).

The marine Mancos Shale underlies and interfingers with sandstones of the Mesaverde Group. The Mesaverde Group is composed of the Castlegate, Sego, and Illes Fork Formations, the latter two of which are separated by coastal-plain coals of the Cameo Coal interval (Fig. 2; Yurewicz et al., 2008; Zhang et al., 2008). The Castlegate, Sego, and Illes Formations include marine sandstones deposited through a series of regressions, separated by tongues of Mancos shale that represent short transgressive cycles. The Williams Fork Formation includes a series of nonmarine shales and disconformable fluvial sandstones (Cumella and Ostby, 2003; Cumella and Scheevel, 2008; Nelson, 2003). Potential source rocks for gas include marine Mancos Shale, coals (including the Cameo coals) within the Illes (northern Piceance) and lower Williams Fork Formations, and nonmarine shale in the Williams Fork Formations (Fig. 2; Yurewicz et al., 2008; Zhang et al., 2008; Harris et al., 2013). With burial and temperature increase, the source rock–generated gas entered the low-permeability sandstones.
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The sandstones of the Mesaverde Group contain subvertical natural opening-mode fractures (Pitman and Sprunt, 1987; Lorenz, 2003; Cumella and Scheevel, 2008; Hooker et al., 2009; Fall et al., 2012). The predominant strike of the fractures is WNW-ESE, locally changing to E-W in the western part of the basin (Pitman and Sprunt, 1987; Lorenz and Finley, 1991; Cumella and Scheevel, 2008; Lorenz, 2003; Baytok and Pranter, 2013). The fractures are confined to the sandstone layers, and do not cross the sand-shale contacts (Pitman and Sprunt, 1987; Cumella and Scheevel, 2003; Lorenz, 2003).

FRACTURE CEMENT PETROGRAPHY

Fractures are mineralized with quartz as the predominant fracture cement in the deeper stratigraphic intervals and with calcite in the shallower parts of the Mesaverde Group (Figs. 3A and 3B; Lorenz and Finley, 1991; Weisenberger et al., 2012). Where both quartz and calcite are present, quartz predates calcite. Several samples contained kaolinite fracture cement (Fig. 3C), and one sample contained barite cement, which postdates calcite. All samples analyzed for this study contained both quartz and calcite (Fig. 3C).

Forty-three samples of completely (27 samples) or partially (16 samples) cemented natural fractures in core from 11 wells in the Piceance Creek and Love Ranch fields, northern Piceance Basin (Fig. 1), were obtained for fluid inclusion analyses. Fracture cement textures were investigated by thin section petrography and SEM-CL imaging.

SEM-CL images were obtained using a Phillips XL30 SEM equipped with an Oxford Instruments MonoCL cathodoluminescence system, operated at 12–15 kV and large sample currents for CL images. Color SEM-CL images were obtained by sequentially acquiring gray-scale images using blue, green, and red filters, compositing these images into red-green-blue (RGB) images, and by combining single RGB images into image mosaics. Contrast and RGB color balance were adjusted using image-processing software.

Quartz Cement

Two types of quartz fracture cement were observed: (1) crack-seal cement bridges that locally span across the fracture aperture, and (2) euhedral cement forming a lining on fracture walls and on earlier-formed cement bridges (Figs. 3A, 4A, and 4B). Cement bridges are surrounded by either fracture porosity or later euhedral cement. SEM-CL imaging revealed
that quartz cement bridges are composed of several tens to hundreds of crack-seal cement layers (Figs. 4B and 4C) indicative of cement precipitation concurrent with fracture opening (Laubach, 2003). Each fracture opening increment was followed by a cement precipitation event that trapped fluid inclusions that comprise a fluid inclusion assemblage, representing a group of fluid inclusions that were all trapped at the same time (Fig. 4B). The petrographic association between crack-seal cement and fluid inclusion assemblages allows interpretation of the fracture cement fluid inclusion record of fluid temperature, pressure, and composition during fracture growth. Euhedral cement growing along the side of crack-seal bridge cement and along fracture walls is generally inclusion-free. Because crack-seal cement increments frequently do not follow a systematic growth sequence across the fracture, we used SEM-CL imaging and textural image interpretation to reconstruct the sequence of crack-seal increments as a prerequisite to correctly interpret the fluid inclusion microthermometric results. This cement sequence was determined for groups of crack-seal increments based on crosscutting relations among different generations of crack-seal cement layers and between crack-seal cement and lateral euhedral cement following criteria summarized by Becker et al. (2010) (Figs. 4C and 4D).

Calcite Cement

Calcite cement is observed either as the only cement present (mostly in fractures in the shallower parts of the Mesaverde Group), or as a secondary cement phase postdating earlier quartz cement. Where present, calcite frequently fills fracture porosity completely (Fig. 3B and 5). Calcite fracture cement occurs as banded columnar calcite with crack-seal texture (Fig. 5A), and as fine- to coarse-crystalline sparry calcite (Figs. 3B and 5B). The sequence of crack-seal increments could not be determined because of ambiguous crosscutting relations of the cement increments. Banded and sparry calcite fracture cement contains both primary and secondary fluid inclusions. Fluid inclusions interpreted as primary relative to the cement growth form clusters or bands within distinct growth layers of the crystal. Inclusions forming bands oriented parallel to the fracture walls are interpreted to have been trapped during repeated crack-seal events of fracture opening and cementation. Inclusions interpreted as secondary occur along sealed microfractures that are oriented along cleavage planes of the calcite and crosscutting growth layers and crack-seal cement increments.

Barite Cement

We observed barite fracture cement forming large subhedral crystals and postdating earlier calcite cement in a single sample at shallower parts of the Mesaverde (2266 m). The barite contained both primary and secondary aqueous and hydrocarbon gas inclusions.

FLUID INCLUSION ANALYSIS

Fluid inclusions in both quartz and calcite cements were analyzed, as well as in one barite crystal. Of 43 fracture samples that were screened, 14 samples contained fluid inclusions large enough to be suitable for microthermometry (8 in quartz, 5 in calcite, 1 in barite).

The observed fluid inclusion assemblages in quartz are composed of two-phase aqueous
liquid-vapor fluid inclusions, most of them coexisting with single-phase hydrocarbon gas inclusions. Several samples of quartz cements were found to contain hydrocarbon gas–rich inclusions only. The coexistence of the hydrocarbon gas and aqueous inclusions within the same fluid inclusion assemblage indicates inclusion trapping in the two-phase immiscible field (Goldstein and Reynolds, 1994), and that the aqueous inclusions are hydrocarbon (CH₄) saturated. Measured homogenization temperatures of the aqueous fluid inclusions thus represent temperatures of inclusion trapping (Goldstein and Reynolds, 1994). Both aqueous and gas inclusions have irregular or rounded shape; the aqueous inclusions contain less than 10 vol% vapor. Most of the inclusions are smaller than 10 µm. Most of the analyzed fluid inclusions in quartz are confined to the center of the crack-seal cement layer and are thus considered primary relative to the age of the hosting crack-seal cement layer. In addition, several fluid inclusion assemblages containing large (~10–30 µm) two-phase liquid-vapor inclusions were observed along microfractures that are perpendicular to the crack-seal layers. Following Becker et al. (2010), who observed similar inclusions in the Travis Peak Formation, we interpret these inclusions as primary or pseudosecondary inclusions that formed along reentrants or deep etches. These inclusions were not included in the analysis of the crack-seal fluid inclusions.

Primary fluid inclusions in calcite and barite form small three-dimensional (3-D) clusters within the crystals. Several trails of secondary fluid inclusions were also observed along sealed microfractures. Calcite cement also contains coexisting two-phase aqueous and single-phase hydrocarbon gas inclusions, or hydrocarbon gas inclusions only, although both types of inclusions are much less abundant compared to quartz fracture cement. Inclusions in calcite are up to ~20 µm in diameter with irregular to negative crystal shapes.

Microthermometric and barometric techniques followed those described in Becker et al. (2010) and Fall et al. (2012). Fluid inclusion microthermometry preceded SEM-CL imaging in order to prevent possible destruction or reequilibration of fluid inclusions due to localized heating from the electron microbeam of the SEM (Xu, 2012). Details are presented in Appendix 1.

**Microthermometry Results**

Homogenization temperatures (Tₒ) of the aqueous fluid inclusions in quartz vary from ~140 °C to ~188 °C (Fig. 6; Table DR1). Homogenization temperatures within individual fluid inclusion assemblages are very consistent and vary by only ~1–4 °C. Many of the observed fluid inclusion assemblages hosted in quartz crack-seal bridges show systematic temperature trends (Fig. 6). For example, homogenization temperatures in a crack-seal quartz bridge of sample 2–12183.5 increase from ~170–175 °C in older cement to ~180–187 °C in the younger cement layers (Figs. 4B and 4D). Because this cement is synkinematic, this temperature trend reflects an increase in temperature during fracture opening. Inclusions in other samples for which the relative timing of crack-seal increments was determined show decreasing temperature trends, while other samples show minor or no temperature trends over time (Fig. 6).

The homogenization temperatures of the fluid inclusions in calcite range from ~135 °C to ~165 °C (Fig. 6; Table DR1 [see footnote 1]), with Tₒ variation of ~1–15 °C within a single fluid inclusion assemblage. These larger temperature variations within individual fluid inclusion assemblages compared to inclusions in quartz are probably caused by partial reequilibration of inclusions in calcite due to the low hardness and cleavage of the host mineral (Goldstein, 1986; Bodnar, 2003). While fluid inclusions are present in crack-seal increments of calcite cements, temperature evolution trends and relative timing of the inclusions could not be determined because of the small number of inclusions in apparently subsequent cement increments.

The homogenization temperatures of the fluid inclusions in barite range from ~115 °C to ~140 °C (Fig. 6; Table DR1 [see footnote 1]). We attribute the relatively large Tₒ variation of up to ~25 °C in some of the fluid inclusion assemblages to be the effect of reequilibration of the inclusions (Bodnar, 2003).

Initial and final ice-melting temperatures were generally difficult to observe due to the small size of the fluid inclusions. However, final ice-melting temperatures were measured for a few large inclusions in most of the studied fluid inclusion assemblages. Final ice-melting

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Footnote 1: GSA Data Repository item 2014268, Fracture cement fluid inclusion results, northern Piceance Basin, Colorado, is available at http://www.geosociety.org/pubs/fr2014.htm or by request to editing@geosociety.org.
ranges of homogenization temperatures (in °C) are given for each fluid inclusion assemblage (FIA). Number of fluid inclusions in each assemblage is given in parentheses. (C) Color scanning electron microscopy–cathodoluminescence (SEM-CL) image mosaic of quartz bridge shown in B. (D) Panchromatic SEM-CL image of C with zones of cement growth highlighted and numbered from oldest to youngest.

temperatures correspond to salinities ranging from 2.0 to 3.0 wt% NaCl equivalent for inclusions in quartz, and from ~2.0 to 3.5 wt% NaCl equivalent for inclusions in calcite and barite (salinity calculation based on Bodnar, 1993). The salinities show no systematic trends with homogenization temperatures across multiple crack-seal increments. The coexistence of aqueous and gas-rich inclusions within single fluid inclusion assemblages indicates the presence of a two-phase aqueous and hydrocarbon gas–saturated fluid.

Microthermometry of the single-phase hydrocarbon inclusions indicates that the fluid in the inclusions is in the liquid state at room temperature, owing to the fact that the fluid inclusions nucleate a vapor bubble upon cooling. The inclusions homogenize to the liquid phase within the range of ~75 °C to ~68 °C. This phase-change behavior indicates the presence of a fluid with density above the critical density of methane (Goldstein and Reynolds, 1994). Homogenization temperatures above ~82.1 °C (the critical temperature of methane) are indicative of components in addition to methane, such as ethane, propane, and/or CO₂, components common in sedimentary systems (Roedder, 1984; Goldstein and Reynolds, 1994). The presence of methane, ethane, and carbon dioxide in single-phase liquid inclusions was confirmed by Raman microspectrometry. Methane, but not ethane and/or CO₂, was also identified in the vapor bubble of aqueous inclusions using Raman microspectrometry.

Geobarometry Results

Aqueous fluid inclusions in quartz fracture cement in the northern Piceance Basin have bulk methane concentrations of ~4000–9000 ppm, corresponding to trapping pressures of ~40–95 MPa (5800–13,780 psi; Fig. 7A; Table DR1 [see footnote 1]). The pressures show no systematic trends for fluid inclusion assemblages trapped in consecutive crack-seal increments of individual quartz cement bridges. Instead, these fluid inclusion assemblages show variations in pressure over time indicative of oscillating pore-fluid pressure conditions during fracture opening and cementation. Similar variable-pressure records were obtained for inclusions in quartz fracture cement in the southern Piceance Basin.
(Fall et al., 2012), and this is to be expected in a dynamic system undergoing continuous fracturing, cementation, and recharge by fluids with variable amounts of gas.

Methane concentrations of fluid inclusions in calcite range from ~3000 ppm to 5500 ppm, corresponding to trapping pressures from ~17 to 60 MPa (Fig. 7A; Table DR1 [see footnote 1]). Methane concentrations of fluid inclusions in barite range from ~3000 ppm to 4000 ppm, corresponding to trapping pressures from ~15 to 52 MPa (Fig. 7A; Table DR1 [see footnote 1]). No systematic pressure trends were observed for inclusions in calcite and barite.

We compared calculated pore-fluid pressures to pressures assuming a lithostatic gradient of 24.8 MPa/km (~1.1 psi/ft), calculated for a depth-averaged rock density of 2.53 g/cm³ using density logs for the southern Piceance Basin (Steve Cumella, 2012, personal commun.), and to pressure for a hydrostatic gradient of 9.9 MPa/km (~0.43 psi/ft) based on a water density of 1.02 g/cm³ (Fig. 7B). The thermobaric lines corresponding to the lithostatic and hydrostatic gradients are plotted in Figure 7A assuming a geothermal gradient of 38.8 °C/km. This paleogeothermal gradient was obtained using fluid inclusion data for two samples collected at different depths in two wells ~6 km (~3.8 mi) apart. Fluid inclusions in crack-seal quartz cement of these two samples (III-11674 and VII-9664) showed relatively constant temperatures across bridges of 186.5–188.8 °C and 156.2–157.8 °C, respectively, indicating that they formed at maximum burial temperature. A well log correlation indicates 92 m of differential uplift of sample VII-9664 relative to sample III-11674, with a corresponding difference in true vertical depth below mean sea level (TVDmsl) between both samples ~790 m. The difference in the temperature at which the two quartz bridges formed thus corresponds to a paleogeothermal gradient of 36.3–41.3 °C/km (mean 38.8 °C/km). This paleogeothermal gradient falls within the range of present-day geothermal gradients of 28 °C/km in the northwestern part of the Piceance Basin and of 43 °C/km in the southeastern part of the Piceance Basin (Johnson and Nuccio, 1986; Zhang et al., 2008).

The calculated fluid inclusion trapping pressures indicate pore-fluid pressures at the time of fracture opening significantly above the hydrostatic pressure gradient (Figs. 7A and 7B). Some of the pressure-temperature points obtained from inclusions in calcite and barite fall below the hydrostatic gradient (Fig. 7B), which we attribute to stretching of inclusions or partial fluid loss after trapping.

**DISCUSSION**

**Timing of Fracture Formation**

The timing for the onset and end of fracture formation can be estimated by combining fluid inclusion microthermometric data with burial history models obtained independently using burial and thermal maturity data, as was previously demonstrated by Hanks et al. (2006), Becker et al. (2010), Duncan et al. (2012), and Fall et al. (2012). Based on data for the base of the Mesaverde Group in the MWX1 well of the Rulison field, Fall et al. (2012) showed that fracture opening in the southern Piceance Basin started at ca. 41 Ma and lasted until 6 Ma. Here, we expand this approach to multiple reservoir depths of the Mesaverde Group by combining the results of this study for the northern Piceance Basin with those of Fall et al. (2012) for the southern Piceance Basin to evaluate the effect of burial depth and temperature on the onset and end of fracture growth in different stratigraphic horizons of the reservoir on a basin-wide scale.

Following Becker et al. (2010) and Fall et al. (2012), we assume that trends in trapping temperature observed across crack-seal cement layers reflect changes in burial temperature and burial depth rather than variation in temperature caused by advective heat transport associated with episodic upward fluid flow. We justify this assumption with the lack of cyclically variable trapping temperatures in the fracture cement record that would be characteristic of rapid upward fluid flow and advective heat transport as inferred for connected fault and fracture systems elsewhere (Eichhubl and Boles, 2000a). Rapid upward fluid flow is necessary to obtain a locally perturbed temperature anomaly (Eichhubl and Boles, 2000a). In a reservoir of low matrix permeability, such flow conditions require a hierarchical network of well-connected fractures.
Fall et al.

Figure 6. Range of trapping temperatures of fluid inclusions in fracture cement and trends of increasing and decreasing temperature plotted against present-day true vertical depth, Piceance Creek and Love Ranch fields. Bars represent range of temperatures for multiple fluid inclusion assemblages or for multiple fluid inclusion assemblages in multiple bridges in any single fracture.

and faults found in some conventional fractured reservoirs (Eichhubl and Boles, 2000b). Such connected fracture systems are generally not characteristic of unconventional reservoirs that require hydraulic fracture stimulation to get a flow response in the wellbore. In the absence of indicators for repeated fluid temperature variations, we relate the observed increasing and decreasing trapping temperature trends in crack-seal cement layers to fracture opening and synkinematic cement growth during prograde burial and exhumation, respectively. The single burial cycle in the Piceance Basin (Fig. 8; Yurewicz et al., 2003, 2008) thus allows an unambiguous correlation between temperature evolution trend observed in the fracture cement inclusion record and time of fracture cement precipitation and fracture opening. Inclusion records from crack-seal fracture cements without temperature change from one fluid inclusion assemblage to the next are interpreted to reflect fracture growth during maximum burial over a period lasting ~25 m.y. (Fig. 8).

Because temperature evolution models are not available for each well we sampled, we adapted the temperature evolution model of the Love Ranch #1 well from Yurewicz et al. (2003, 2008) for the sampled wells in the Piceance Creek and Love Ranch fields (Fig. 8A). Temperature evolution curves were calculated for each sample by correcting measured depth (MD) to TVDMSL and using the paleogeothermal gradient of 38.8 °C/km obtained from fluid inclusion records in samples III-11674, and VII9664 (Fig. 8A). Along these temperature evolution curves, the observed range of fluid inclusion trapping temperatures for each fracture is highlighted using bold lines (continuous for quartz; dashed for calcite and barite) in Figure 8A. This correlation indicates that the fractures in the northern Piceance Basin formed between 39 and 6 Ma.

A similar plot (Fig. 8B) was calculated based on the burial temperature profile of the MWX1 well and using the microthermometric data of Fall et al. (2012) for 12 samples from four wells in the southern Piceance Basin (Fall et al., 2012, model 1 in their fig. 11) and using a geothermal gradient of 43 °C/km (Johnson and Nuccio, 1986; Zhang et al., 2008). According to this assessment, fractures in the southern Piceance Basin formed between 41 and 6 Ma.

Fluid inclusion trapping temperatures in calcite and barite cements are generally lower than those in quartz cement (Fig. 7), suggesting that calcite precipitated at shallower depths than quartz, consistent with our petrographic observations that calcite precipitated after quartz (Fig. 3C). Calcite fracture cement thus formed during maximum burial and early exhumation of the reservoir.

For determination of the timing of calcite fracture cementation, we only considered fluid inclusions in synkinematic crack-seal calcite. The larger $T_h$ variation within fluid inclusion assemblages in calcite (up to ~15 °C) compared to $T_h$ variations in quartz (~4 °C) suggests that some of the inclusions in calcite were stretched subsequent to trapping, resulting in homogenization temperatures that are elevated compared to inclusions unaffected by stretching (Goldstein and Reynolds, 1994; Bodnar, 2003).
Stretching might also have contributed to the relatively large homogenization temperature ranges observed in barite (up to 25 °C). To avoid overestimating the timing of fracture opening of calcite- and barite-filled fractures based on fluid inclusion temperatures that may have been artificially elevated, we plotted the range of minimum $T_h$ values obtained in each individual fluid inclusion assemblage at any sampled depth in Figure 8A, and we assumed that these minimum temperatures most closely represent the actual (prestretching) temperatures of the inclusions.

While the temperature record of individual fractures lasted for generally less than 10 m.y., the overall duration of fracture opening is

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Figure 7. (A) Trapping temperatures plotted against calculated trapping pressures of fluid inclusions in quartz, calcite, and barite fracture cements from the Piceance Creek and Love Ranch fields. The diagram shows isopleths from 2500 to 9000 ppm in the H$_2$O-NaCl-CH$_4$ system for the 2 wt% NaCl-CH$_4$ pseudobinary system, calculated using the equation of state of Duan and Mao (2006). Also shown are the thermobaric gradients of 24.8 MPa/38.8 °C and 9.9 MPa/38.8 °C, corresponding to the lithostatic and hydrostatic gradients, respectively. (B) Calculated trapping pressures plotted against inferred maximum burial depth, including an estimated 1500 m (~5000 ft) of Tertiary section removed by erosion in the last 10 m.y. in the Piceance Basin. Lithostatic and hydrostatic gradients are as in A. Data points below the hydrostatic gradient in both figures are probably the result of stretching of the fluid inclusions to produce elevated (incorrect) temperatures. FI—fluid inclusion; p (after sample depth)—primary fluid inclusions; s—secondary fluid inclusions.
Although our sample record may not completely capture the onset of fracture opening in a particular stratigraphic horizon, we consider our sample record sufficiently robust to discern a progression in the onset of fracture opening from deeper to shallower horizons, a depth interval of ~1550 m, over ~15–20 m.y. in the northern Piceance Basin. A similar progression over ~10 m.y. is observed in the southern Piceance Basin over a depth interval of ~690 m. Our results for the southern Piceance Basin suggest that fracture formation starts when the formation reaches a burial temperature of ~140–150 °C and ceases when the burial temperature falls below ~150–160 °C. In the northern Piceance Basin, the onset of fracture formation correlates with temperatures of 170 °C in deeper stratigraphic levels, and with ~140 °C in shallower horizons.

No clear time difference between shallower and deeper stratigraphic horizons is observed for the termination of fracture opening. Fracture opening appears to have ceased concurrently for all stratigraphic horizons at around 6 Ma in both the northern and southern Piceance Basin.

Mechanism of Fracture Formation

A comparison of the timing of the onset of fracture formation at 39 Ma inferred here for the northern Piceance Basin with models of gas and oil generation of Yurewicz et al. (2008) and Zhang et al. (2008) suggests that fracture opening and elevated pore-fluid pressures generally coincided with gas generation, consistent with similar findings by Fall et al. (2012) for the southern Piceance Basin. The general overlap in timing of fracture formation and gas generation, in combination with the ubiquitous presence of gas-rich fluid inclusions and with high pore-fluid pressures calculated from fluid inclusion methane concentrations, is consistent with, though not necessarily indicative of, natural hydraulic fracturing driven by gas generation. The onset of fracture formation 20–30 m.y. before exhumation excludes the possibility that the onset of fracture formation was triggered by stress changes associated with exhumation. The deep burial conditions at the onset of fracture formation also exclude high pore pressures generated by compaction disequilibrium as the trigger for fracture formation because the sandstone was largely cemented at this stage of burial (Ozkan, 2010; Ozkan et al., 2011; Stroker et al., 2013).

We cannot exclude the possibility that pore-fluid pressures were already elevated because of earlier compaction disequilibrium at the time when significant methane generation began. However, the lack of fractures that predate methane charge seems to indicate that compaction disequilibrium did not drive fracture formation. Further-
more, the observed fluctuation in paleo-pore-fluid pressure requires a process of overpressure generation that was active during fracture formation as opposed to a mechanism of overpressure generation that predated fracture formation.

Three gas sources have been considered for the Mesaverde Group: (1) the marine Mancos Shale that underlies and interferes with sandstones of the lower Mesaverde Group; (2) coal layers, including the Cameo coal, and (3) nonmarine shale (Fig. 2; Johnson and Rice, 1990; Johnson and Roberts, 2003; Yurewicz et al., 2008; Zhang et al., 2008; Harris et al., 2013). Hydrous pyrolysis experiments and kinetic yield models by Yurewicz et al. (2008) suggested that the coal layers provided a more prolific source of gas compared to both the marine and nonmarine shales (Figs. 9A and 9B), whereas compositional analyses of gas produced from the Mesaverde Group in the southern Piceance Basin are consistent with gas sourced dominantly from marine shale (Johnson and Rice, 1990; Harris et al., 2013).

Kinetic yield models of Yurewicz et al. (2008) for the Love Ranch #1 well indicate that gas generation started at around 55 Ma, with peak generation at 40 Ma for the marine shale source and at around 30 Ma for coal (Figs. 9A and 9B), leveling off at around 15 Ma. If we assume that fracture formation in the lower Mesaverde Group was the response to gas generation in both marine shale and coals, we find that fracture formation initiated at ~30%–35% of total cumulative gas generated from these two sources (Figs. 9A and 9B). Fracture formation concurrent with gas generation is consistent with fracturing driven by gas generation in the source rock layers and subsequent or concurrent gas migration into adjacent low-permeability reservoir layers. This process may have resulted in cycles of episodic pore-fluid pressure buildup, subsequent fracture opening, pressure decay, and fracture cementation resulting in the formation of crack-seal fracture cement textures. Alternatively, episodic opening of fractures may reflect a temporal and spatial shift in opening displacement among adjacent fractures within a layer held under constant pore-fluid pressure. Such lateral variability in fracture opening in a single layer may account for the limited duration of opening of a single fracture compared to the overall duration of fracture opening in these layers.

Fracture opening in the Piceance Basin did not commence until 41 Ma, or ~15 m.y. after the beginning of gas generation (Fig. 9). The time interval between the onset of gas generation and the onset of fracturing can be explained by the time necessary to build up pore-fluid pressures to levels high enough to overcome the least principal stress in order to fracture the rock. It has previously been suggested that early-generated gas might have been lost by updip migration along the laterally continuous sandstone layers of the Sego and Castlegate Formations in the northern Piceance Basin and of the Iles Formation in the southern Piceance Basin (Johnson and Nuccio, 1986; Johnson, 1989; Johnson and Roberts, 2003; Yurewicz et al., 2008), thus delaying a pore pressure buildup. Numerical simulations of quartz pore cementation in Williams Fork sandstone by Ozkan (2010) for a well in the Mamm Creek field, southern Piceance Basin, suggested that present-day matrix porosity values were reached between 55 and 35 Ma, allowing the possibility that the delay in pore-fluid pressure increase relative to gas generation can be attributed to a decrease in matrix porosity and permeability during prograde burial. The onset of fracture growth would thus be controlled by the permeability evolution of the matrix in addition to gas generation and charge. Yurewicz et al. (2008) attributed the delay in gas charge to the absorption of early-generated gas onto coals of the Mesaverde Group.

Lorenz and Finley (1991) proposed that late Eocene–Oligocene (30–36 Ma) top-to-the-west–directed high-angle thrusting of the White River uplift reactivated earlier fractures in the Mesaverde Group. Based on our results, this is also the time period when fractures in the Mesaverde formed. However, the time progression in the onset of fracture formation from deeper to shallower stratigraphic horizons within the Mesaverde Group appears to be inconsistent with fracture formation in response to such tectonic processes. Instead, we assume that stress changes associated with these tectonic processes would have affected different stratigraphic horizons of the Mesaverde Group at the same time, resulting in a synchronous onset of fracture formation throughout the section. The onset of fracture formation at approximately the same time in both the northern and southern Piceance Basin and in all fields that we analyzed, without systematic regional differences, is inconsistent with fracture formation triggered by reservoir-scale tectonic or structural processes. We therefore conclude that the fluid inclusion record is most consistent with incipient fracture formation driven by gas.

Figure 9. Cumulative gas yields (billion cubic feet/km³) for main source intervals of marine shales within the Mancos Shale, coastal-plain coals (including the Cameo coals), marine shale and coal combined, and nonmarine shales within the Iles and Williams Fork Formations (A) at Love Ranch #1 and (B) MWX1 (after Yurewicz et al., 2008). Dashed lines indicate onset and end of fracture opening for the entire reservoir interval as shown in Figures 8A and 8B. Onset of fracture formation in the lower Mesaverde Group approximately coincides with 30%–35% of total cumulative gas generated from combined marine shale and coal sources.
Implications for Gas Migration

The observed progression of fracture formation from deeper to shallower stratigraphic horizons provides constraints on the rate and distance of gas migration during the onset and termination of fracture growth. We tested multiple hypotheses of gas generation, gas migration, and timing of fracture growth, represented schematically in Figure 10 for two layers (L1 and L2) that approximate deeper and shallower reservoir horizons of the Mesaverde Group. We consider scenarios with and without upward migration of gas from L1 into L2. Rapid upward gas migration from deeper sources such as the Mancos Shale and coals in the Iles Formation to the stratigraphically higher intervals of the Williams Fork Formation, a vertical distance of 1550 m in the northern Piceance Basin and 690 m in the southern Piceance Basin, could have occurred over 15–20 m.y. in the northern and over 10 m.y. in the southern Piceance Basin. Such slow rates of upward gas migration of ~70–80 m/m.y. in the southern Piceance Basin, could have occurred over 15–20 m.y. in the northern and over 10 m.y. in the southern Piceance Basin. Such slow rates of upward gas migration are consistent with the observed progression in fracture onset between L1 and L2. We speculate that such rapid upward migration would have left a fracture record between 40 and 30 Ma in the shallower horizons of the Mesaverde Group, which is not observed (Figs. 8A and 8B). Instead, the observed delay is consistent with fracture formation in response to gas generation in close proximity to the fractured reservoir layers, either without large-scale migration (Figs. 10A, 10E, and 20F) or with only slow gas migration (Figs. 10C and 10D). Slow gas migration from the Mancos Shale and coals in the Iles Formation to the stratigraphically higher intervals of the Williams Fork Formation, a vertical distance of 1550 m in the northern Piceance Basin and 690 m in the southern Piceance Basin, could have occurred over 15–20 m.y. in the northern and over 10 m.y. in the southern Piceance Basin. Such slow rates of upward gas migration could also account for the late fracture opening at temperatures <150 °C observed in the Love Ranch field (wells 4 and 5 in Fig. 8A), well below the 170 °C associated with fracturing onset in the deeper Iles Formation. However, if the slow rate of upward gas migration persisted throughout the time of gas charge, slow migration would delay not only the onset of fracture formation, but also the end of fracture formation between shallow and deep reservoirs (Figs. 10C and 10D), especially if gas migration was slower than the rate of exhumation (Fig. 10C). Such a delay in the end of fracture formation is not observed (Figs. 8A and 8B). The timing and, in particular, the onset of fracture formation (Figs. 8A and 8B) are most consistent with fracture formation driven by gas that was locally sourced, without significant upward gas migration, from source layers undergoing gas generation at different isotherms (Figs. 10E and 10F). These findings are similar to our interpretation of fracture growth in the Piceance Basin.

Figure 10. Models for fracture formation in tight-gas reservoirs by gas generation and resulting pore-fluid pressure increase in two stacked layers, L1 and L2. (A) Gas generation in interbedded source/reservoir layers without upward gas migration. Onset of fracture formation is coincident with the source/reservoir layers crossing the depth or isotherm of 30% of total gas generation (TGG30). Fracture formation is assumed to start under similar temperature and burial depth conditions in both layers, with the onset of fracturing in L2 delayed relative to L1. (B) Fracture formation concurrent in L1 and L2 coincident with source layers interbedded with L1 reaching TGG30 and gas migrating rapidly from L1 into L2. L2 is not required to enter the gas maturation window for fractures to form in L2. (C) Same as B but with gas migrating at slow rate from L1 to L2, thus delaying onset of fracture formation in L2 relative to L1. Gas migration rate is assumed to be slower than rate of subsidence. (D) Same as C but with rate of gas migration faster than rate of subsidence. (E) L2 reaching TGG30 at lower temperature and at shallower depth than L1 because of differences in organic matter composition or differences in time-temperature history between L1 and L2 (e.g., owing to a change in burial rate for L1). Onset of fracture formation is delayed relative to L1 but to a lesser extent compared to A. Unlike B, C, and D, no upward gas migration is assumed. (F) Same as E, but TGG30 in L2 is reached at greater depth/higher temperature compared to L1. Delay in fracturing onset in L2 is longer compared to A.
Natural hydraulic fracturing of tight-gas sandstone reservoirs

The structural and diagenetic attributes of fractures observed in the Mesaverde Group are characteristic of tight gas- and oil-producing sandstone reservoirs in general (Laubach et al., 2004a, 2004b; Hanks et al., 2006; Becker et al., 2010). Based on the similarity in fracture attributes, we infer that slow fluid-driven hydraulic fracture growth induced by thermocatalytic gas generation is a widespread phenomenon that can occur in tectonically quiescent basins undergoing prograde burial diagenesis. In the absence of tectonic processes, these fractures may lack the connectivity to larger faults, however, and thus be of little consequence to regional cross-stratigraphic flow. Under suitable tectonic conditions, such fractures may form the nucleus that propagates and coalesces with adjacent fractures into connected fracture systems and fault zones that serve as formation-scale flow conduits (Aydin, 2000; Eichhubl and Boles, 2000b).

CONCLUSIONS

Fluid inclusion analyses of crack-seal fracture cement, precipitated concurrently with fracture opening, provide a record of temperature, fluid compositional, and fluid pressure conditions throughout the time of fracture opening in Mesaverde tight-gas sandstone reservoirs. This record constrains the timing of fracture formation within different stratigraphic horizons of the reservoir and provides insight into the interplay among gas generation, gas migration, and fracture formation.

Coexisting aqueous and hydrocarbon gas inclusions indicate that fractures opened during gas charge of the reservoir. Systematic changes in fluid inclusion homogenization and trapping temperatures, increasing from ~170°C to ~188°C and then decreasing to ~140°C, reflect fracture formation during prograde burial to near maximum burial conditions and during early exhumation of the reservoir, from 41 Ma to 6 Ma in the northern Piceance Basin, and from 39 Ma to 6 Ma in the southern Piceance Basin. The onset of fracture formation 20–30 m.y. before exhumation excludes the possibility that fracture formation was triggered by stress changes associated with exhumation. Calculated pore-fluid pressures of 40–95 MPa (5800–13,780 psi) during fracture formation indicate significant pore-fluid overpressures consistent with fracture formation driven by gas generation. An observed progression in the onset of fracture formation from deeper and shallower reservoirs of the Mesaverde Group over a period of 15–20 m.y. in the northern and over a 10 m.y. period in the southern Piceance Basin is inconsistent with fracture formation triggered by stress changes in response to regional structural or tectonic processes. This age progression is also inconsistent with rapid upward gas migration from deeper source layers into shallower reservoirs, which we expect would have resulted in the synchronous onset of fracture formation throughout the producing formation interval. Instead, we propose that fracture formation was driven by gas generation and migration into adjacent reservoir layers, and the concomitant pore-fluid pressure increase led to natural hydraulic fracturing. This largely thermally activated process occurred earlier in deeper reservoir horizons and later in shallower horizons. Gas generation thus developed a pervasive fracture network that permitted continuous gas accumulation in Mesaverde reservoirs.

APPENDIX 1

Fluid Inclusion Microthermometry

Microthermometric analysis of fluid inclusions was carried out using a FLUID, Inc.–adapted, U.S. Geological Survey–grade low-temperature stage mounted on an Olympus BX 51 microscope equipped with a 40× objective (numerical aperture = 0.55) and 15× oculars. The stage was calibrated using the CO$_2$ ice-melting temperature at ~56.6 °C of H$_2$O–CO$_2$ synthetic fluid inclusions, and the ice-melting temperature at 0 °C and critical homogenization temperature at 374.1 °C of pure H$_2$O synthetic fluid inclusion standards (Sterner and Bodnar, 1984). Liquid-vapor homogenization temperatures ($T_h$) were determined to ±0.05°C by thermal cycling using temperature steps of 0.1 °C (Goldstein and Reynolds, 1994). Final ice-melting temperatures were also determined to ±0.1 °C by the same thermal cyclic technique.

Fluid Inclusion Geobarometry

The methane concentration of the aqueous fluid inclusions can be used to calculate fluid pressure during fracture opening and cementation (Pironon et al., 2003; Becker et al., 2010; Fall et al., 2012). We followed the technique of Lin et al. (2007) to determine the pressure within aqueous fluid inclusions at room temperature using Raman spectroscopic analyses. Raman analyses were carried out using a JY Horiba LabRam HR (800 nm) spectrometer at Virginia Tech. The settings of the spectrometer were optimized to obtain the best peak intensity and are identical to those described in detail by Lin et al. (2007) and Becker et al. (2010). The method of estimating the pressures in the two-phase aqueous fluid inclusions at room temperature is based on the CH$_4$ Raman symmetric stretching ($\nu_1$) peak position with the laser beam focused on the vapor bubble of the inclusions. The position of the CH$_4$ Raman peak is density dependent and is related to the vapor bubble pressure as described in detail by Lin et al. (2007). The fluid pressure at inclusion trapping conditions was calculated following Becker et al. (2010) using the equation of state of Duan and Mao (2006), the vapor bubble pressure at room temperature, the homogenization temperature, and the salinity of the aqueous fluid. This calculation assumes that the fluid is adequately characterized by the H$_2$O–NaCl–CH$_4$ system, and that the vapor phase is pure CH$_4$.

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