PATTERNS OF MONTOYA GROUP DEPOSITION, DIAGENESIS, AND RESERVOIR DEVELOPMENT IN THE PERMIAN BASIN

Rebecca H. Jones

ABSTRACT

Rocks composing both the Montoya (Upper Ordovician) and Fusselman (Lower Silurian) Formations were deposited during the global climate transition from greenhouse conditions to unusually short-lived icehouse conditions on a broad, shallow-water platform. The Montoya and the Fusselman also share many reservoir characteristics and have historically been grouped together in terms of production and plays. Recently, however, the Montoya has garnered attention on its own, with new gas production in the Permian Basin and increased interest in global Ordovician climate. Recent outcrop work has yielded new lithologic and biostratigraphic constraints and an interpretation of four third-order Montoya sequences within Sloss's second-order Tippecanoe I sequence.

The Montoya Group comprises the Upham, Aleman, and Cutter Formations, from oldest to youngest. The Upham contains a basal, irregularly present sandstone member called the Cable Canyon. The boundary between the Montoya and the Fusselman is readily definable where a thin shale called the Sylvan is present but can be difficult to discern where the Sylvan is absent. Montoya rocks were deposited from the latest Chatfieldian to the end of the Richmondian stage of the late Mohawkian and Cincinnatian series (North American) of the Upper Ordovician.

Montoya reservoir quality is generally better in the northern part of the Permian Basin where it is primarily dolomite compared to limestone Montoya reservoirs in the south. Reservoir quality is also better in the lower part of the unit compared to the upper, owing to a predominance of porous and permeable subtidal ooid grainstones and skeletal packstones in the former and peritidal facies in the latter.

INTRODUCTION

The Montoya Group comprises a moderately thick (100 to 600 ft) Upper Ordovician carbonate ramp succession present in both outcrop and the subsurface of

West Texas and southeastern New Mexico. The four Montoya Group formations, the Upham (and Cable Canyon Member sandstone), Aleman, and Cutter have been defined and well-studied in outcrop but are generally not correlated to the subsurface. Montoya Group thickness reaches a maximum of 590 ft thick in outcrop (Pope, 2004a) and 600 ft thick in the subsurface in Loving, Pecos, Ward, and Winkler Counties and as the Montoya-equivalent Maravillas Formation in Brewster County (Texas Water Development Board, 1972). The subsurface distribution limit is reached in Garza, Borden, Howard, Glasscock, and Reagan counties to the east, Culberson and Jeff David counties to the west and Hockley and Lynn Counties to the north (Figure 1) (Texas Water Development Board, 1972). In southeastern New Mexico, the Montoya's presence extends to Chaves and Roosevelt Counties in the north and Dona Ana County in the west (Wright, 1979). The Montoya Group was largely deposited on the Middle-Upper Ordovician Simpson Group but locally overlies on the Lower Ordovician Ellenburger or equivalent. The Sylvan Shale, where present, and the Fusselman Formation generally overlie the Montoya.

Montoya reservoirs are better known for their recent gas production than their relatively low cumulative oil production. From 1993- 2007, 497 BCF of gas and 16.6 MMbbl of oil were produced from Montoya reservoirs (Drilling Info, 2007). Notable gas fields include Block 16 (109.9 BCF) and R.O.C. (27.8 BCF) in Ward County and Waha (40.0 BCF) in Pecos/Reeves Counties (all amounts produced from 1993-2007), and Beall (31.3 BCF produced from 1999-2007) in Ward County (Texas Railroad Commission, 2008). Top producing oil fields with production clearly attributed to reservoirs developed in Montoya rocks (and cumulative production as of the year 2000) include Abell field in Pecos and Crane Counties, Texas (12.6 MMbbl), Tex-Hamon in Dawson County, Texas (4.8 MMbbl), Halley (3.0 MMbbl) and Monahans North (1.0 MMbbl) fields in Winkler County, Texas, and Justis field in Lea County, New Mexico (11.0 MMbbl) (Dutton and others, 2005) (Figure 2).

Other Texas fields (and counties) with Montoya production include Martin (Andrews); East Tank (Borden); Abell Northeast (Crane); Tippett North and Tippett West (Crockett); Effort (Dawson); TXL (Ector); Azalea East (Midland); Abell West, GMW, Heiner, Lehn-Apco, Lehn-Apco North, Mesa Vista, Oates Southwest, Pecos

Valley, Pecos Valley East, Pecos Valley South, and Pikes Peak (Pecos); Worsham-Bayer (Reeves); McEntire, WAM, WAM South, and Westbrook (Sterling); Tokio (Terry); Beall East, and Halley South, and Wink South (Ward).

Distinguishing oil production from rocks in the Montoya Group vs. that from rocks in the Fusselman Formation in the Permian Basin is difficult due to the practice of reporting Montoya and Fusselman production together, the lack of seal between the Montoya and Fusselman, and potential commingling with production from the Ellenberger Formation in places where there is an unconformable contact between Montoya and Ellenburger rocks.

Outcrops studies from the mountains of West Texas and New Mexico and in the Marathon Region in southwestern Texas describe the Montoya Group as a series of subtidal carbonate facies deposited in inner- to outer-ramp settings during waning greenhouse conditions. Very little has been published on the subsurface. This report synthesizes previous work and describes new core and outcrop data with the aim of improving the understanding of Montoya reservoirs and their relationship to outcrops in West Texas and southeastern New Mexico.

PREVIOUS WORK

General accounts of the Montoya reservoir were included in early publications by Jones (1953), Herald (1957), Galley (1958), Howe (1959), and Pratt and Jones (1961). Outcrop descriptions were also published early, by Pray (1958) and Pratt and Jones (1961), and have continued with publications by McBride (1970), Measures (1984, 1985a, and 1985b), and Brimberry (1991). Pope and Steffen (2003) and Pope (2002a, 2002b, 2004a, 2004b, and 2004c) recently developed a sequence stratigraphic model based on outcrop observations and related Montoya facies to regional climatic events. Several authors have described cores taken during recent Montoya exploration: Ball (2002 and 2003) and Behnken (2003) described a core from Dollarhide field and Thomas and Liu (2003) presented observations from cores in a study area including Ward, Pecos, and Reeves Counties.

REGIONAL SETTING

Paleogeography and Climate

In the Late Ordovician, the relatively stable conditions that had prevailed for most of the Ordovicianbegan to change. Landmasses were assembled into a supercontinent, Gondwana, and three major terranes; North America was the Laurentian terrane (Cocks and Torsvik, 2004). Within Laurentia, present-day West Texas and southeastern New Mexico were located near 30ºS (Blakey, 2004) (Figure 3). Gondwana began to migrate across the South Pole in the Late Ordovician — a move that likely caused a unique, short-lived episode of glaciation during this waning period of greenhouse conditions (Crowley and Baum, 1995, Pope, 2004b).

Based on Webby (2004), the Montoya Group was deposited from about 452-448 Ma in a mature passive margin setting characterized by fluctuating climatic conditions. The nearest highlands were located in northern New Mexico (Figure 3). The subtidal, gently dipping ramp carbonates making up the Montoya formed during the transitional period to an unusual, short-lived Gondwana glaciation within a longer period of overall greenhouse conditions characterized by high CO_2 concentrations (Pope, 2004b). Global sea-level was at or near the Paleozoic maximum and an extensive oceanic upwelling zone along the southern margin of Laurentia, in what now is New Mexico, Texas, and Oklahoma, resulted in deposition of subtidal ramp carbonates contining up to 70% spiculitic chert by volume and 1 to 5 weight percent phosphate (Pope and Steffen, 2003). Faunal assemblages suggest that a deep marine basin occupied the area basinward of the Ouachita-Marathon overthrust (Figure 4). Glaciation of the region reached a maximum during the Hirnantian Stage (Figure 5), following deposition of the Montoya (Saltzman and Young, 2005; Young and others, 2005). Greenhouse conditions would again prevail by the end of the Silurian.

Isotopic Evidence for Climate Change

Carbon isotope stratigraphy of K-bentonite-bound horizons, biostratigraphy, and facies analysis has been used to identify the onset of oceanic upwelling that was associated with cooling and glaciation. Upwelling of cooler nutrient-rich waters (Si and PO_4), increased primary productivity and resulted in preferential sequestration of isotopically light (^{12}C) carbon during the Hirnantian.. (Young and others, 2005). Following a period of upwelling and carbon sequestration, disproportionately more heavy carbon (^{13}C) was sequestered in the seawater causing fewer nutrients to be available for carbonate production. Seawater with higher $\delta^{13}C$ ratios circulated onto the carbonate platform and became incorporated into skeletal packstones and grainstones, resulting in a distinct isotopic enrichment in these skeletal-rich strata compared to the mud- and chert-rich strata below.

Two isotope excursions have been documented in the Late Ordovician: the first occurred at during the early Chatfieldian stage and has been referred to as the Guttenberg carbon isotope excursion (GICE); the second occurred in the Hirnantian. Both events have been associated with glaciation of Gondwana and tied to changes in ocean circulation. During the GICE, $\delta^{13}C$ ratios were enriched by ~3‰ in Upper Ordovician strata from numerous locations throughout North America, including the Viola Group in the Arbuckle Mountains of Oklahoma, equivalent intervals in Kentucky, Virginia, and West Virginia (Young and others, 2005), and the Nashville Dome area in Tennessee (Holland and Patzkowsky, 1997). The GICE marks a fundamental change in the style of carbonate deposition, the cause of which has been interpreted as a minor episode of Gondwana glaciation. These locations also have similar ε_{Nd} ratios, indicating that they are from the same continuous body of water (Holmden and others, 1998), eliminating the possibility that the $\delta^{13}C$ excursions are due to geochemically-distinct epicontinental masses of water, rather than climatic changes. The Hirnantian isotope excursion included enrichment of both $\delta^{13}C$ and $\delta^{18}O$ by ~2‰ in brachiopod samples from around the world, suggesting a short-lived period of global glaciation lasting approximately 0.5-1 million years (Brenchley and others, 1994).

Thus, the Montoya Group was deposited between two short-lived episodes of continental glaciation during an overall greenhouse climate. Most Montoya facies reflect the upwelling and sea-level rise associated with the transitional conditions preceding Hirnantian glaciation; however, the siliciclastic basal Cable Canyon Member was likely deposited during lowstand conditions that prevailed immediately following the first, minor glaciation event. Evidence for this conclusion comes from recent work on the Eureka Quartzite in Nevada, which is approximately time equivalent to the Cable Canyon in West Texas and southeastern New Mexico outcrops (Pope and Steffen, 2003). Eureka Quartzite workers concluded that the GICE was followed by a significant fall in sea level with deposition of lowstand clastics as a result of continental glaciation (Saltzman and Young, 2005).

Structure

A cratonic origin has been invoked by both historical and contemporary authors for Cambrian – Ordovician clastics, including the Middle-Late Ordovician Simpson Group. Exposed Precambrian basement and Cambrian granitic plutons formed paleotopographic highs (Pedernal Uplift, Diablo Arch) that sourced the Cambrian Bliss Sandstone (Goldhammer and others, 1993). The Pedernal Massif in central and north-central New Mexico was also a regional high and supplied sediment for Simpson sandstones and silts in southeastern New Mexico (Kottlowski, 1970) and likely the southeastern extension of these deposits into West Texas. By ca. 450 Ma, the cratonic sediment supply no longer reached the area south of Ouachita-Marathon fold belt, according to neodymium isotope analysis of Maravillas Formation sediments from this area (Gleason and others, 1995). Prior to this time, siliciclastics were transported from eroding highs in the northwest and deposited as the Cable Canyon Member sandstone. The lack of local highs in this shallow marine platform setting eliminates the possibility of a local provenance for these siliciclastics (Figure 4). An isopach map of the Cable Canyon sandstone in southeastern New Mexico (Figure 6) lends further support to this idea by depicting a northwest-southeast-striking locus of deposition and thickening towards the northwest.

Post-depositional structural thickening of both Montoya and Simpson Group rocks in front of the Ouachita-Marathon overthrust (Reeves, Pecos, and Ward Counties) was observed in 3-D seismic (Hardage, 1999), reprocessed 2-D seismic (Swift and others, 1994), and well data from this area, with repeated section created through both high-angle reverse faults (Figure 7) and overturned structures (Figure 8). Not only are these observations relevant to wireline correlations, but they also may explain the discrepancy between modern and historical interpretations of depositional environment; without seismic data, the structurally thickened strata may have misled previous workers into interpreting a basin setting for the Montoya Group, e.g., figure 11 in Galley, 1958. Deeper water conditions were likely present basinward of the Ouachita-Marathon overthrust (Figure 3) where the Maravillas Formation is present, but not landward, where the Montoya facies are characteristic of a shallow carbonate platform setting.

FACIES AND SEDIMENTOLOGY OF THE MONTOYA GROUP

The Montoya Group was initially described as a formation with two members, the Second Value and the Par Value, before being renamed as a group with four subdivisions: the Cable Canyon Sandstone, the Upham Dolomite, the Aleman Formation, and the Cutter Formation (Kottlowski et al., 1956). The Cable Canyon is now referred to as a member of the Upham Formation in both outcrop and the subsurface, (Pope 2004a and Thomas and Lui, 2003) (Figure 5). Montoya Group equivalents include the Maravillas Formation in the Marathon region of Texas and the Viola Group in Oklahoma and northern Texas (Anadarko Basin).

Unconformities are present both above and below the Montoya Group. Conodont data (Sweet, 1979) indicate major breaks in sedimentation at both boundaries; however, the basal unconformity has not been observed in Oklahoma (Dennison, 1997). The Montoya Group was deposited on Simpson Group carbonates and sandstones in the center of the Montoya subcrop area in West Texas and southeastern New Mexico (Figure 1). Where the Simpson is absent, i.e., in the eastern and the very northern Midland Basin, western Delaware Basin, and far western Texas and New Mexico outcrops (Figure 1), the Montoya overlies the Ellenburger or the equivalent El Paso Group. The upper unconformity marks a significant period of erosion related to post-depositional uplift (Mears and Dufurrena, 1984); in some cases large portions of the upper Montoya Group were removed . Montoya rocks reach a maximum of 590 ft of thickness in outcrop (Pope, 2004a) and over 600 ft in subcrop (Wright, 1979). Outcrops are present in the Beach, Hueco, and Franklin Ranges of West Texas and the Sacramento, San Andres, Franklin, and Caballo Ranges of southeastern New Mexico (Pope, 2004a). Montoya outcrops have also been reported in the Baylor and Sierra Diablo Ranges in Texas (Jones, 1953).

The Montoya is Cincinnatian series in age, having been deposited during the Edenian and Richmondian stages of the Upper Ordovician, based on conodont biostratigraphy (Sweet, 1979) and ages assigned to conodont species zones (Webby, 2004) (Figure 5). The Viola Group in Oklahoma has been often discussed as equivalent; however, conodont data from outcrops in this region (Derby and others, 1991) indicate that only the Upper Viola Springs Formation, the Welling Member, and the Sylvan Formation are truly age-equivalent (Figure 5). Even more problematic are correlations with the Maravillas Formation in the Marathon Uplift region of West Texas. Graptolite biostratigraphy (Goldman and others, 1995) and graptolite-conodont age equivalents (Webby, 2004) indicate that these strata are Richmondian and therefore only overlap with the latter half of Montoya deposition (part of the Aleman through Cutter deposits) in Texas/New Mexico and latest Sylvan deposits in Oklahoma.

The Montoya Group was largely deposited in a shallow-water platform setting characterized by normal marine conditions. Cool water currents from both the north and south were present along the western coast of Laurentia (Figure 3) related to the pending Hirnantian glaciation. The southerly ocean currents resulted in upwelling of cool waters in present day West Texas and New Mexico with deposition of cherty carbonate updip and cherty shale downdip as observed in outcrop (Pope, 2004a). These chert trends are also interpreted to be present in the subsurface of the northwestern and southwestern Permian Basin, respectively (Figure 4).

Upham Formation and Cable Canyon Member

The Upham Formation, including the Cable Canyon Member sandstone where present, rests unconformably on the karsted surface of the Lower Ordovician El Paso Group in most outcrops and on the Simpson or Ellenburger in subcrop and in outcrops east of the Hueco Range. The Cable Canyon member is thin (10 cm or less) and irregularly present in the Franklin Mountains, ranges from less than 0.5 m to over 2 m in the Sacramento Mountains (Brimberry, 1991), and is greater than 15 m thick in the Cooks Range (Pope, 2002) (outcrop locations shown in Figure 1). It is poorly documented in the subsurface, but similarly thin (2 to 20 ft) (Thomas and Liu, 2003). The Cable Canyon Member and Upham Formation are both exposed at the Scenic Drive and McKelligon

Canyon outcrops in the Franklin Mountains on the northern edge of El Paso, Texas (Figure 9).

The Cable Canyon and Upham are interpreted to have been deposited from 452-451 Ma during the Edenian stage, a period of time corresponding to the late *confluens* and *velicuspis* conodont zones. These units comprise one third-order sequence and document the start of marine transgression following a significant hiatus in deposition (Figures 5 and 10). Cable Canyon siliciclastics were likely deposited during lowstand and then reworked during transgression.

Facies

Cable Canyon lithofacies include gravel conglomerate and carbonate-cemented (primarily dolomite) quartz sandstone (Pope, 2004a; Bruno and Chafetz, 1988) that are poorly sorted with grains ranging from 0.1 mm to >2.0 mm in outcrop (Brimberry, 1991). This unit is dominantly medium-grained but coarsens where more thickly deposited, with a grain-size profile that increases from the base to middle and then decreases from the middle to top of the unit in New Mexico outcrops (Bruno and Chafetz, 1988). This carbonate-rich siliclastic unit was originally deposited by traction transport, i.e. fluvial or aeolian, processes (Bruno and Chafetz, 1988), but most of the original cross-bedding has been masked by extensive burrows, which can include 1.5 m deep vertical *Skolithos* burrows that are filled with quartz sandstone (Pope, 2002b). Quartz grains are well-rounded; the plentiful fossil fragments include crinoids, gastropods, brachiopods, and bryozoans (Pope, 2002b). At Scenic Drive outcrops in the Franklin Mountains, the Cable Canyon/Upham contact is gradational, with sand incorporated into the lowermost Upham (Figure 11a). This contact is much sharper at McKelligon Canyon, where the Cable Canyon comprises sandstone with thin lenses of carbonate, which appear to have been reworked from the underlying El Paso Group (Figure 11b). In subsurface Permian Basin cores, the Cable Canyon consists of poorly sorted, well-rounded, variable coarseness sandstone and sandy packstone with skeletal fragments (Thomas and Liu, 2003).

The Upham comprises coarse-grained skeletal wackestones-packstones and grainstones that are variably colored, massive, and can be highly bioturbated in outcrop (Pope, 2004a; Pope and Steffen, 2003). This basal Upham can contain up to 30% quartz at the very base in places where a distinct Cable Canyon unit is absent. Faunal assemblages include corals, crinoids, brachiopods, bryozoans, gastropods, receptaculitid algae, and nautiloids. The dominant skeletal wackestones-packstones are punctuated and capped by coarse-grained crinoidal grainstone beds and a massive unit with rare cross-bedding, respectively. Phosphate (pellets and replacement of bryozoans) and hardgrounds were also observed in outcrop (Pope, 2002b).

Bioturbated skeletal wackestones containing large coral (Figure 12) were observed at the Scenic Drive outcrop and color variation was observed at the McKelligon Canyon outcrop (Figure 13), both in the Franklin Mountains. These dolostones and locally present limestones contain phosphate (pellets, encrusted hard grounds, and replaced skeletal grains) and chert (irregular nodules, diagenetic replacement) (Pope and Steffen, 2003). The fauna at Scenic Drive include a distinct species of solitary rugose coral, *neotryplasma floweri*, that are known only to exist in this area and the Ural region of Russia (Elias, 1986). In subsurface cores from the Permian Basin, the lower Upham comprises dark-colored chert bearing skeletal packstones, wackestones, and mudstones and the upper Upham comprises light-colored packstones and grainstones with a coarser texture and more diverse fauna (Thomas and Liu, 2003).

Depositional Setting

The Cable Canyon Member, Upham Formation, and the lower part of the Aleman Formation represent inner-, mid-, and outer-ramp facies within a second-order transgressive systems tract (Figure 14a). The Cable Canyon was deposited in waters 5-15 m deep as a sand-wave complex deposited by asymmetrical tidal currents (Bruno and Chafetz, 1988) during initial sea-level rise and may represent reworked siliciclastics from earlier traction deposits (Bruno and Chafetz, 1988) or sand dune deposits (Pope, 2002b) deposited during lowstand following a very brief, pre-Montoya episode of glaciation described earlier. The source of siliciclastics was likely eroding Precambrian basement highs to the northwest, a source also invoked for the Middle Ordovician Simpson Group. The Cable Canyon isopach map over southern New Mexico (Figure 6) lends support to this idea, in that the locus of deposition trends northwest-southeast with thickening towards the interpreted sediment source in the northwest. This thickness variation is interpreted to be purely depositional, rather than evidence for erosion because the contact between the Cable Canyon and the overlying Upham Formation is gradational (Bruno and Chafetz, 1988). Contacts observed in the Franklin Mountains were also gradational and characterized by high sand content in the lower Upham (Figure 11). The Upham burrowed skeletal wackestones-packstones were deposited in the shallow subtidal mid-ramp with warm waters, that developed during continued sea-level rise within the same 3rd-order sequence (Pope, 2004c, Figure 10). Energy levels increased and shoals likely developed, in which the crinoid-rich grainstones were deposited. The hardgrounds and phosphate and iron coatings were likely created subaqueously under anoxic conditions, when frequent sea-level rises and upwelling currents brought phosphate-rich waters into this dominantly shallow ramp (Pope, 2002b).

Aleman Formation

The Aleman Formation overlies the Upham Formation in West Texas and New Mexico outcrops (outcrop locations shown in Figure 1) and is exposed at the McKelligon Canyon outcrops in the Franklin Mountains on the northern edge of El Paso, Texas (Figure 9). Several described cores have been assigned to the Aleman formation, including cores from the southern Delaware Basin (Thomas and Liu, 2003) and a recent core from Dollarhide field in Andrews County (Ball, 2002 and Behnken, 2003), which was also examined in this study and will be discussed in the Reservoir Geology section. The Aleman Formation is Maysvillian to Richmondian in stage, corresponds to the *robustus* and early *grandis* conodont zones, and was deposited from 451-449.5 Ma. Portions of the Aleman are contained within two 3rd-order sequences (Figures 5 and 10)

Facies

The Aleman Formation comprises interbedded carbonate and chert. The carbonate has been extensively dolomitized with the exception of a locally present basal limestone. In outcrop, chert is abundant (30-40%) and phosphate content is similar to that of the Upham Formation (Pope and Steffen, 2003).

A thin-bedded chert interval (Figure 15), overlain by a middle grain-rich interval (Figure 16), and an upper nodular chert interval (Figure 17) were observed at outcrops in

McKelligon Canyon in the Franklin Mountains (Figure 9). These patterns have been observed in other West Texas and New Mexico outcrops, e.g., Pope 2002a, 2002b, 2004a, 2004b, 2004c, and Pope and Steffen, 2003. These workers describe three facies in the Aleman: 1) even-bedded laminated calcisiltite or mudstone and spiculitic chert, 2) skeletal wackestones to packstones with discontinuous bedded to nodular chert, and 3) skeletal packstone to grainstone with abundant crinoids, bryozoans and brachiopods interbedded with thin coral bafflestones. The lower Aleman is dominantly facies 1 with some overlying facies 2, the middle Aleman is facies 3, and the upper Aleman is facies 2.

Three types of chert have been interpreted from these outcrops by Pope (2004a): primary, early diagenetic and late diagenetic. Primary chert was deposited as thin beds or lenses of sponge spicules, between layers of mudstone and calcisiltite. The lack of sedimentary structure suggests that the spicules were deposited below storm wave base. Most chert nodules observed in outcrop were surrounded by bent laminations, suggesting that they formed on the seafloor before complete lithification and therefore represent an early stage of diagenesis. Relict sponge spicules were also observed within chert nodules in Aleman Formation outcrops in the Silver City Range in southwestern New Mexico (Geeslin and Chafetz, 1982). Late diagenetic chert formed through three mechanisms: 1) as replacement of evaporate nodules in tidal flat facies, 2) replacement of evaporates in subtidal facies, which were likely formed by burial brines during reexposure of platform, or 3) veins or tabular beds.

<u>Depositional Setting</u>

The lower bedded chert facies in Aleman Formation were deposited in a deep ramp setting characterized by cool waters and rare storm waves. Sponge spicules were likely transported into this setting from up ramp and interbedded with the *in situ* calcisiltite and mudstone (Pope, 2002b). The middle Aleman skeletal packstone to grainstone facies was deposited in a warm-water high-energy shoal, as evidenced by cross-bedding (Pope, 2002b). Both the lower and upper Aleman contain skeletal wackestones to packstones with bedded and nodular chert, which are representative of a slightly shallower setting between the deep ramp calcisiltite and grainstone shoals. The chert breccia facies rarely encountered in the Aleman represents slumping of early-formed chert (Pope, 2002b).

Cutter Formation

The Cutter Formation overlies the Aleman Formation in West Texas and New Mexico outcrops (outcrop locations shown in Figure 1) and is exposed at the McKelligon Canyon outcrops in the Franklin Mountains on the northern edge of El Paso, Texas (Figure 9). It is eroded in some places in the subsurface, at least partially owing to post-depositional structural uplift (Mears and Dufurrena, 1984). The Cutter Formation is Richmondian stage, corresponds to the *grandis* conodont zone, and was deposited from 450-448 Ma. It comprises one full and one partial third-order sequence (Figures 5 and 10).

Facies

Bioturbated skeletal wackestones and laminated mudstones, evaporates, and rare secondary silica nodules (evaporate replacement) comprise the Cutter Formation (Pope and Steffen, 2003). Distinct facies include skeletal packstones (bryozoans, brachiopods, and crinoids abundant), burrowed mudstone with locally interbedded green-brown shale, and laminated and fenestral mudstone (Pope, 2002). This overall light-colored fine-grained interval consists of dolomite with minor chert (Pope, 2004a). Brachiopod wackestone with lenses of crinoidal packstone (Figure 18), overlain by wackestone to laminated mudstone (Figure 19) were observed at outcrops in McKelligon Canyon in the Franklin Mountains (Figure 9). In core from the subsurface Permian Basin, dark-colored chert-bearing wackestones and mudstones of the Lower Cutter are overlain by packstones and grainstones with decreasing chert content (Thomas and Liu, 2003).

Depositional Setting

These facies are interpreted to represent shallow subtidal to peritidal deposition. Skeletal packstones were deposited during a relatively brief period of open marine conditions; burrowed mudstones were deposited in a restricted subtidal setting (lagoon);

and laminated and fenestral mudstones were deposited in a tidal flat setting, with semi-arid and humid climates, respectively (Pope, 2002b).

The Viola Group and Sylvan Formation in Oklahoma

The Viola Group and Sylvan Shale are approximate Montoya Group equivalents in eastern Texas, Oklahoma, and Arkansas. Age-equivalency based on the latest conodont biostratigraphy and age data (Figure 5) show that the lower part of this group, the Lower Viola Springs Formation, does not have age-equivalent Montoya formations, and that the upper part of the group, the Upper Viola Springs Formation, including the Welling Formation, is equivalent to the Cable Canyon Member and Upham and lower Aleman Formations in the Montoya Group. The Sylvan Shale was deposited at the same time as the upper Aleman and Cutter Formations. Duration of deposition of the Sylvan has been estimated at 3 million years and there is no evidence of an unconformity at its base (Dennison, 1997), but age relationships suggest that a significant hiatus occurred between deposition of the Sylvan and the overlying Keel Formation (Figure 5).

Facies

The lower Viola Springs Formation comprises interbedded laminated calcisiltite or carbonate mudstone and bedded and nodular chert in the Arbuckle Mountains (Mitchell, 2003). This is overlain by bioturbated thinly-bedded calcisiltite and mudstone with nodular chert. Skeletal wackestone-packstone with chert nodules and medium to thick bedding characterize the upper Viola Springs Formation and skeletal packstones and grainstones with thick bedding characterize the Welling Formation (Mitchell, 2003). Primary porosity is present in the grainstones and closely spaced post-depositional (Pennsylvanian) fractures are present in the mud-rich rocks of the Viola Springs (Dennison, 1997). The consistent thickness (100-300 ft) and clay-richness of the Sylvan create an effective seal for the highly productive Viola Group (Dennison, 1997).

Depositional Setting

The depositional setting of the Viola Group has been interpreted to be similar to that of the Montoya (Mitchell, 2003). The group comprises an overall shallowing-upward succession with deeper water mudstones grading into shallow water grainstones (Dennison, 1997). Contourite and turbidite sedimentary structures suggest that the carbonate ramp was steep (Pope, 2002b). The Sylvan shale was deposited in a shallow subtidal marine setting with low energy (Sternbach, 1984) when a new sediment source of clay abruptly ended carbonate deposition (Dennison, 1997).

Sylvan Formation in Texas

The Sylvan Formation in Texas is not age equivalent to that in Oklahoma (Figure 5) and therefore may represent entirely different shale. The Sylvan nomenclature has been applied to an irregularly present thin shale has been used as a high gamma-ray wireline log pick to separate the Montoya Group from the overlying Fusselman in the subsurface. There can be numerous high gamma-ray responses in the upper portions of the Montoya (Figure 20) that do not represent shales, as we discovered when logging the upper part of the Montoya in a core from Dollarhide field in Andrews County, Texas. Distinct Montoya and Fusselman facies were recognizable in core, but the portion of core that would have contained the formation boundary was missing, so any shale/gamma ray relationships could not be confirmed by this core. Nonetheless, the lack of correlation between shale and high gamma ray wireline response in the upper Montoya suggest that caution should be used in picking the Montoya/Fusselman boundary on the basis of high gamma-ray wireline log responses alone.

The Maravillas Formation

The Maravillas outcrops in the Marathon Uplift area of southwestern Texas are considered equivalent to the Montoya (McBride, 1970) although biostratigraphy indicates both a significant hiatus during the early period of Montoya deposition and an overall deeper depositional environment setting for a distinct biofacies when compared to the Montoya Group and equivalents in the rest of North America (Goldman and others, 1995). This formation describes facies deposited in the area in West Texas labeled "deep marine basin" in Figure 4. Neodymium isotope analysis of sediments from this area suggests that sediments in this area were not derived from the Laurentian craton; rather, passive margin shales with strongly negative ε_{Nd} values gave way to less negative ε_{Nd}

orogenic turbidites from the emerging Appalachian orogen at this time (Gleason and others, 1995).

The Maravillas is a 60-500 ft thick chert-rich formation with three informal members based on variations in lithology (McBride, 1970). The lowermost member contains dominantly limestone with chert, the middle member contains dominantly chert with limestone, and the thin upper member contains chert and shale. The upper member has been previously called the Solitario and the Persimmon Gap Members. The upper shale is likely correlative with the Sylvan in West Texas and New Mexico where present and in southeastern Oklahoma (Wilson, 1954).

The Maravillas comprises 40% black chert, 30% limestone (calcarenite, micrite, and marlstone), 14% shale, 10% non-black chert, 5% limestone pebble conglomerate, and 1% dolomite (McBride, 1970). Bedding is regular with thickness varying from three to 12 inches. Whereas some earlier authors invoke a shallow-marine setting invoked on the basis of bryozoans and primary chert, McBride (1970) suggests that the depositional setting of these rocks was deep-water slope to basin floor and concludes that the bryozoans were transported and the chert was secondary. Additional evidence for a deep-water depositional environment include the lack of typical shallow water structures, such as wave formed ripples and bedding, coupled with the presence of characteristic deep water features, including slump structures, coarse conglomerates, and anoxic conditions indicated by high organic matter and lack of bioturbation (McBride, 1970).

Sequence Stratigraphy of the Montoya Group

The Upper Ordovician is part of the Tippecanoe I second-order supersequence set (Sloss, 1988) of the Tippecanoe first-order megasequence (Sloss, 1963) (Figure 5). As described early, the Montoya Group was deposited between two short-lived episodes of Gondwana glaciation and therefore most facies were strongly influenced by the transitional greenhouse-icehouse climate. Sea level changes were therefore higher amplitude (20-50 m) and more frequent than would be expected during normal greenhouse conditions (Read and others, 1995). Montoya deposition following a significant mid-Tippecanoe I hiatus after deposition of the Middle-Upper Ordovician Simpson Group and was fully deposited before the beginning of the Tippecanoe II second-order sequence in the Silurian (Figure 5). This depositional hiatus occurred throughout the study area and is interpreted to have had a particularly long duration in the Marathon Uplift area (Goldman and others, 1995). Montoya facies are interpreted to have been deposited during a 2nd-order highstand systems tract (HST) and transgressive systems tract (TST) (Figure 14). The Montoya Group comprises four complete 3rd-order sequences and portions of two others (Figure 10).

<u>First and Second-Order Sequences</u>

The Montoya Group was deposited during the transgressive leg of the Tippecanoe first-order megasequence (Sloss, 1963), called the Tippecanoe I second-order supersequence set (Sloss, 1988). Within this second-order sequence, skeletal sandstone and granule conglomerate were deposited in an inner ramp setting (likely reworked lowstand deposits) followed by mid-ramp transgressive systems tract packstones and grainstones, and then deep ramp calcisiltite and spiculitic chert (Figure 14a) Continued deep ramp deposition continued followed by chert-bearing wackestones and packstones as conditions shallowed to mid-ramp. Then, packstones and grainstones were deposited followed by burrowed skeletal wackestones and laminated and fenestral mudstones during the highstand systems tract (Figure 14b). A simple link can be made between second-order systems tracts and formation names: the Upham was deposited during initial transgression across the ramp, the Aleman during major deepening (late TST/early HST), and the Cutter during widespread highstand peritidal conditions (Pope and Steffen, 2003).

<u>Third-Order Sequences</u>

The four Montoya Group lithostratigraphic formations can be fit into four widespread, plus two irregularly present, third-order sequences (1 to 3 m.y.) (Figure 10). Workers in the subsurface Permian Basin have defined four third-order sequences in the Montoya Group: the first sequence comprises the Cable Canyon member (lowstand deposits) and the Upham (transgressive and highstand systems); the second sequence comprises the lower part of the Aleman shallowing-upward succession; the third sequence comprises the upper part of the Aleman shallowing-upward succession; and the fourth sequence comprises the Cutter Formation (Thomas and Liu, 2003). The north-south outcrop section scheme developed by Pope (2004a) (Figures 1 and 10) places the Cable Canyon and Upham facies in the initial transgressive sequence, the lower Aleman cherty facies and the lower part of the medial subtidal grain-rich Aleman facies in the second sequence, the upper part of the medial subtidal grain-rich Aleman facies, the upper Aleman cherty facies, and part of the Cutter peritidal facies in the third sequence, and the Cutter peritidal facies in the fourth sequence. An additional sequence of Cable Canyon and Upham facies is present locally at the base and an additional sequence of shallow subtidal mid-ramp carbonates with open marine fauna is present locally at the top.

Reservoir Geology

In cores from a study area including Ward, Pecos, and Reeves Counties, facies consists of dark-colored chert-bearing wackestones and mudstones overlain by chert-free packstones with a grainstone cap, overlain by numerous coarsening-upward cycles of chert-bearing packstone to grainstone (Thomas and Liu, 2003). These facies were interpreted to correspond to the lower, middle, and upper Aleman, respectively.

A core at Dollarhide field in Andrews County (from the Dollarhide 25 2-S well) has also been assigned to the Aleman Formation by Ball (2002 and 2003) and Behnken (2003). This core was also examined by this study, but correlation to the outcrop formations not made. Incomplete coring of the Montoya interval (including no coverage of the Montoya/Fusselman boundary) and known unconformities at both the base and top of the Montoya were factors in deciding not to attempt these correlations without additional data.

Facies observed in the Dollarhide core (Figure 20 and 21) include chert mudstone (Figure 22), mudstone, dolowackestone (Figure 23), and dolopackstone – grain-dominated dolopackstone (Figures 24 and 25). A section of chert-bearing mudstone is present from 8457-8479 ft; chert is also present in wackestones at the base of the core. Chert nodules contain relict sponge spicules and fracturing and microporosity are developed around their rims (Figure 26). Interpretation of depositional environments from this core has been difficult. The lack of diagnostic exposure surfaces coupled with abundant chert and numerous thin grain-rich intervals make definitive assignment of facies to a peritidal vs. subtidal environment challenging. This base of the core is clearly subtidal, with upward-shallowing cycles comprised of peloidal packstone or wackestone (Figures 24 and 23) at cycle bases and grain-dominated dolopackstones (Figure 25) at cycle tops; however, the upper part of the core has been interpreted as both peritidal ((Ball, 2002 and Behnken, 2003) and restricted subtidal to transitional (this study). We were not convinced that there were sufficient exposure surfaces or diagnostic peritidal features, such as development of fenestral porosity, to definitively place this core in a peritidal setting. We also were not convinced that an interpretation of karst (Behnken, 2003) could be supported by this core. The limited areas of intensely fractured strata could represent local deformational features, similar to those observed in outcrop. Additional studies of nearby cores to provide context will be necessary to resolve the ambiguity about the depositional environment of this core.

Further compounding the problem of subsurface-outcrop correlations is the lack of distinct wireline log characteristics. Wireline log correlations are generally problematic in the Montoya, particularly in differentiating the Montoya from the overlying Fusselman. Both units have low gamma ray responses and wireline porosity is often related to dolomitization, rather than facies. Examination of picks from a database provided by Geological Data Services and published wireline correlations suggests that the combined thickness of the Fusselman and Montoya is often simply halved to make a top Montoya pick where the Sylvan shale is absent. The wireline data give little if any indication of facies, so it is difficult to get away from this approach; however, it is far from ideal. Isopach maps constructed from such picks show neither the true deposition thickness nor the magnitude of unconformities.

Whole core porosity and permeability data from the Dollarhide core show that the best reservoir facies are grain-dominated dolopackstones, located in the lower third of the core (Figures 20 and 25). Porosity ranges from 7-13.2% and permeability from 1.5-183 md in these fabrics. Porosity ranges from 0.3-16.1% and permeability ranges from 0.01-183 md permeability (k90) throughout the cored interval. Porosity and permeability are highest in dolomitized rocks, making mineralogy prediction important; however, predicting mineralogy with grain density data alone can be misleading in this core. Thin sections demonstrate that many intervals contain a mix of dolomite and chert with grain densities ranging from 2.71- 2.79 g/cc. This range reflects a mix of dolomite grain density of 2.85 g/cc and chert (quartz) grain density of 2.65 g/cc (Klein, 1993). Limestones composed of calcite with a grain density of 2.71 g/cc (Klein, 1993) and a small amount of dolomite could also fall into this range.

Reservoir Development

Many Montoya reservoirs have been recently developed for gas production, which has now far surpassed production from the limited number of oil reservoirs. As of 2007, 497 BCF of gas and 16.6 MMbbl of oil have been cumulatively produced from Montoya reservoirs (Drilling Info, 2007). Notable gas fields include Block 16 (109.9 BCF) and R.O.C. (27.8 BCF) in Ward County and Waha (40.0 BCF) in Pecos/Reeves Counties (all amounts produced from 1993-2007), and Beall (31.3 BCF produced from 1999-2007) in Ward County (Texas Railroad Commission, 2008). Top producing oil fields with production clearly attributed to reservoirs developed in Montoya rocks (and cumulative production as of the year 2000) include Abell field in Pecos and Crane Counties, Texas (12.6 MMbbl), Tex-Hamon in Dawson County, Texas (4.8 MMbbl), Halley (3.0 MMbbl) and Monahans North (1.0 MMbbl) fields in Winkler County, Texas, and Justis field in Lea County, New Mexico (11.0 MMbbl) (Dutton and others, 2005) (Figure 2).

Reservoir Distribution

The Montoya is thickest in Pecos, Reeves, and Ward Counties, Counties, as shown by an isopach generated from a database of picks supplied by Geological Data Services (Figure 27). This is a structurally complex area, as mentioned earlier, and thicknesses are probably not representative of deposition, but rather may reflect repeat section through high angle reverse faulting and overturned structures (see Figures 7 and 8). Representative maximum depositional thicknesses are present in part of Ward, Loving, Winkler, and Culberson counties. The group thins quickly to the east (becoming absent in Borden Howard, Glasscock, and Reagan Counties) but oversteps the underlying Simpson group to the west (extending as far as Otero County, New Mexico) (Galley, 1958). Post-depositional erosion due to structural uplift is at least in part responsible for this thinning; up to 17% of the original section has been removed in some areas (Mears and Dufurrena, 1984).

Oil reservoirs are developed predominantly in the dolomitized subtidal facies (skeletal grain-dominated packstones and packstones) located in the northern part of the Permian Basin; gas reservoirs are developed in areas with sufficiently hydrocarbon maturity to yield gas. In both cases, traps are structural. Commingling with the overlying Fusselman is common, but the Sylvan shale, where present, and tight peritidal facies of Cutter Formation can sufficient seal the reservoir interval, as exemplified by Dollarhide field, where the Fusselman reservoir has watered out but the Montoya produces (Ball, 2003).

Porosity Development

Porosity development in the Montoya Group is controlled by both facies and diagenesis. The highest porosity has been developed in dolostones, which are more abundant in the northern part of the Permian Basin, whereas limestones are dominant in the Marathon region outcrops and southern Permian Basin. The transition from limestone to dolostone occurs in Hudspeth, Culberson, Reeves, Andrews, Martin, and Howard Counties and in the Franklin Mountain outcrops (Jones, 1953). Porosity is also better developed in subtidal ramp facies of the lower Montoya (average 6.2%) than in the dominantly peritidal facies of the upper Montoya (average 2.5%) at Dollarhide field in Andrews County (Behnken, 2003). The highest reservoir quality occurs in the lower part of the reservoir where dolomitized ooid grainstones and skeletal packstones have both moldic and intercrystalline porosity; however, lower quality chert-bearing dolomudstones with intercrystalline and fenestral porosity in the upper part of the reservoir are also productive (Behnken, 2003). Chert intervals are also productive at Waha field (Reeves County), where moldic spicules, small pores (< 0.05 mm pore throats), microporosity in fine-grained skeletal grainstone with chert, and slightly dolomitized skeletal packstone with chert constitute part of the Aleman pay zone (Thomas and Liu, 2003).

Coarse dolomite with intercrystalline porosity was observed in thin sections taken from dolopackstones and grain-dominated dolopackstones in the Dollarhide core described in this study (Figures 24 and 25). These samples are likely represent of the highest unfractured reservoir properties, with whole core porosity of 8.1 and 8.4% and permeability of 40.9 and 6.33 md, respectively. Higher porosity and permeability values were observed in chert mudstones but likely represent fracture, not matrix properties. Some microporosity was observed around chert nodule rims (Figure 26), but its contribution to reservoir porosity is relatively minor.

Pore lining and poikilotopic dolomite and tabular to acicular anhydrite have reduced reservoir porosity, as has late calcite, which is present in oomolds and fractures at Dollarhide field (Behnken, 2003). By contrast, caverns and fractures have enhanced porosity and permeability in some Montoya reservoirs (Gibson, 1965). Natural fractures are a key component of porosity and permeability development in horizontal wells producing from the Viola Group in southern Oklahoma (Candelaria and Roux, 1997).

Traps, Seals, and Sources

Most Montoya trap include structural and/or fault closure. The trap at Dollarhide field is a fault-bounded anticline structural trap (Figure 21); however, stratigraphic trapping may occur through changes from subtidal to peritidal facies upsection and dolomitization and several karsting events (Ball, 2002)..A lack of effective barrier (shale or porosity change) between the Montoya and the overlying Fusselman in many areas (for example, where the Sylvan is absent) has allowed hydrocarbons to migrate upwards into the Fusselman (Wright, 1979) and in some cases the Montoya reservoir is connected to the underlying Ellenburger (Gibson, 1965). Given the continuous Montoya/Fusselman oil column observed in many fields, it seems likely that Montoya would share the Fusselman oil source, which has been clearly identified as the Upper Devonian Woodford shale (Williams, 1977).

Opportunities for Additional Resource Recovery

The location of currently producing gas fields overlaps with the area of greatest thickness and deepest burial, as well as greatest structural complexity. Additional gas resources may be recoverable from the southern Delaware Basin with detailed 3-D seismic interpretation. This will be necessary to understand the complex structural traps and repeat section characteristic of the area (Figures 7 and 8).

Careful mapping of facies and mineralogy may also lead to identification of bypassed pay. As shown in the model developed from outcrop (Figure 10), the mid-ramp subtidal facies (skeletal grain-dominated packstones with the highest reservoir quality) are thickest in medial positions on a landward-basinward transect. Placing subsurface data in terms of this model would allow for identification of the mid-ramp facies fairway and thereby the best locations for recompletions or new drilling. The outer ramp chert mudstones and wackestones are lower reservoir quality, but may also hold bypassed resources in areas where the grain-dominated packstones have been produced or waterflooded with the appropriate reservoir management strategies.

SUMMARY AND CONCLUSIONS

The Montoya Group of the Permian Basin reflects a unique transitional climate, during which greenhouse conditions were changing to reflect the pending glaciation of Gondwana, which would occur immediately following deposition. This changing climate had a profound effect on sea-level fluctuations and facies, including 1) high-amplitude, frequent sea-level changes (four complete and two partial third-order sequences), 2) carbonate depositional environments ranging from peritidal to outer ramp, and 3) an abundance of chert and phosphate from upwelling waters. Montoya oil reservoirs are dominantly developed in mid-ramp skeletal grain-dominated packstones, particularly in the northern part of the Permian Basin, where porosity has been enhanced through dolomitization. Gas production has superseded oil, in terms of both quantity and recent interest, and is focused in the southern part of the Delaware Basin, where reservoir quality is likely lower due to more distal facies, but hydrocarbon maturity obviously more advanced. A deep marine equivalent, the Maravillas Formation, is present in the Marathon Uplift area but not known to be productive. The Viola Group of Oklahoma is also considered an equivalent, although facies/age relationships do not exactly match.

Extensive work in Montoya Group outcrops in West Texas and New Mexico has resulted in a sequence stratigraphic context for the formations. The limited core examined in this report suggests that outcrop models can be applied to the subsurface;

however, additional core work is needed to fully establish this relationship and utilize these models for reservoir development. Further rock-based will also be necessary to establish the reservoir architecture, facies patterns, and porosity/permeability relationships required for recovery of the remaining oil and gas resources in Montoya reservoirs.

REFERENCES

Adler, F.L., 1971, Future petroleum provinces of the Mid-Continent: Tulsa Geological Society Digest, v. 38, p. 985-1042.

Ball, B.C., 2003, Identifying bypassed pay in the Fusselman and Montoya reservoirs of the Dollarhide field, Andrews County, Texas, *in* Hunt, T.J., and Lufholm, P.H. (eds.), The Permian Basin: Back to Basics: West Texas Geological Society Publication 03-112,p. 1–12.

Ball, B.C., 2002, A Fusselman and Montoya core from the Dollarhide field, Andrews County, Texas, *in* Hunt, T.J., and Lufholm, P.H. (eds.), Permian Basin: Preserving our Past—Securing our Future: West Texas Geological Society Publication 02-111, p. 207–208.

Behnken, F.H., 2003, Montoya conventional core description, depositional lithofacies, diagenesis and thin section petrography from the Pure Resources, Inc., Dollarhide Unit 25-2-S, Andrews County, Texas, *in* Hunt, T.J., and Lufholm, P.H. (eds.), The Permian Basin: Back to Basics, West Texas Geological Society Publication 03-112, p. 13–35.

Bergström, S.M. and Mitchell, C.E., 1986, The graptolite correlation of the North American Upper Ordovician Standard: Lethaia, v. 19, no. 3, p. 247-266.

Blakey, R.C., 2004, Regional Paleogeographic Views of Earth History, Sedimentation, Tectonics, and Paleogeography of the North Atlantic Region, Late Ordovician: http://jan.ucc.nau.edu/~rcb7/450NAt.jpg.

Brenchley , P.J., Marshall, J.D., Carden, G.A.F., Robertson, D.B.R., Meidla, T., Hints, L., and Anderson, T.F., 1994, Bathymetric and isotopic evidence for a short-lived Late Ordovician glaciation in a greenhouse period: Geology, v. 22, p. 295-298.

Brimberry, D.L., 1991, Depositional and diagenetic history of the Late Ordovician Montoya Group, Sacramento Mountains, south-central New Mexico, *in* Johnson, K.S. (ed.), Late Cambrian-Ordovician geology of the southern Midcontinent, 1989 symposium: Circular - Oklahoma Geological Survey, 1991, v. 92, p. 154-170.

Bruno, L. And Chafetz, H.S., 1988, Depositional environment of the Cable Canyon Sandstone: A Mid-Ordovician sandwave complex from southern New Mexico: New Mexico Geological Society Guidebook, 39th Field Conference, Southwestern New Mexico, p. 127-134.

Candelaria, M.P., and Roux, B.P., 1997, Reservoir analysis of a horizontal-well completion in the Viola Limestone ‘‘chocolate brown zone,’’ Marietta Basin, Oklahoma, *in* Johnson, K.S. (ed.), Simpson and Viola Groups in the southern Midcontinent, 1994 Symposium: Oklahoma Geological Survey Circular 99, p. 183–193.

Chenoweth, P.A., 1966, Viola oil and gas fields of the Mid-Continent: Tulsa Geological Society Digest, v. 34, p. 110-118.

Cocks, L.R. and Torsvik, T.H., 2004, Major terranes in the Ordovician, *in* Webby, B.D., Paris, F., Droser, M.L., and Percival, I.G. (eds.), The Great Ordovician Biodiversification Event, Columbia University Press, p. 61-67.

Collins, E.W., and Raney, J.A., 2000, Geologic Map of West Hueco Bolson, El Paso Region, Texas: Bureau of Economic Geology Miscellaneous Map 40, The University of Texas at Austin, Austin, Texas.

Crowley, T.J. and Baum, S.K., 1995, Reconciling Late Ordovician (440 Ma) glaciation with very high (14X) CO_2 levels: Journal of Geophysical Research, v. 100, no. D1, p. 1093-1101.

Derby, J.R., Bauer, J.A., Miller, M.A., Creath, W.B., Repetski, J.E., Dresbach, R.I., Ethington, R.L., Loch, J.D., Stitt, J.H., Sweet, W.C., McHargue. T.R., Taylor, J.F., Miller, J.F., and Williams, M., 1991, Biostratigraphy of the Timbered Hills, Arbuckle, and Simpson Groups, Cambrian and Ordovician, Oklahoma: A review of the correlation tools and techniques available to the explorationist: Oklahoma Geological Survey Circular 92, p. 15-41.

Denison, R.E., 1997, Contrasting sedimentation inside and outside the southern Oklahoma aulacogen during the middle and late Ordovician, *in* Johnson, K.S. (ed.), Simpson and Viola groups in the southern Midcontinent, 1994 Symposium: Oklahoma Geological Survey Circular 99, p.39-47.

DrillingInfo, 2007, Montoya production data accessed from http://www.drillinginfo.com on July 10, 2007.

Dutton, S. P., Kim, E. M., Broadhead, R. F., Breton, C. L., Raatz, W. D., Ruppel, S. C., and Kerans, Charles, 2005, Play analysis and digital portfolio of major oil reservoirs in the Permian Basin: The University of Texas at Austin, Bureau of Economic Geology Report of Investigations No. 271, 287 p., CD-ROM.

Elias, R.J., 1986, New Late Ordovician solitary rugose coral with perforate septa: Journal of Paleontology, v. 60, no. 1, p. 14-25.

Flower, R.H., 1969, Early Paleozoic of New Mexico and the El Paso region: Guidebook - El Paso Geological Society, Annual Field Trip, Issue 3, Ordovician symposium, p. 31-101.

Frenzel, H.N., Bloomer, R.R., Cline, R.B., Cys, J.M., Galley, J.E., Gibson, W.R., Hills, J.M., King, W.E., Seager, W.R., Kottlowski, F.E., Thompson, S., III, Luff, G.C., Pearson, B.T., and Van Siclen, D.C., 1988, The Permian Basin region, *in* Sloss, L.L. (ed.), Sedimentary cover—North American Craton.: The Geology of North America, Geological Society of America, Boulder, Colorado, v. D-2, p. 261–306.

Galley, J.D., 1958, Oil and gas geology in the Permian Basin in Texas and New Mexico, *in* Weeks, L. G. (ed.), Habitat of oil—a symposium, Tulsa, Oklahoma: American Association of Petroleum Geologists, p. 395–446.

Geeslin, J.H. and Chafetz, H.S., 1982, Ordovician Aleman ribbon cherts: An example of silicification prior to carbonate lithification: Journal of Sedimentary Petrology, v. 52, no. 4, p. 1283-1293.

Gibson, G.R., 1965, Oil and gas in southwestern region—geologic framework, *in* Young, A., and Galley, J.E. (eds.), Fluids in the subsurface environment: American Association of Petroleum Geologists Memoir 4, p. 66–100.

Gleason, J.D., Patchett, P.J., Dickinson, W.R., and Ruiz, J., 1995, Nd isotopic constraints on sediments sources of the Ouchita-Marathon fold belt: Geological Society of America Bulletin, v. 107, no. 10, p. 1192-1210.

Goldhammer, R.K., Lehman, P.J., and Dunn, P.A., 1993, The origin of high-frequency platform carbonate cycles and third-order sequences (Lower Ordovician El Paso GP, West Texas): Constraints from outcrop data and stratigraphic modeling: Journal of Sedimentary Petrology, v. 63, no. 3, p. 318-359.

Goldman, D., Bergström, S.M., and Mitchell, C.E., 1995, Revision of the Zone 13 graptolite biostratigraphy in the Marathon, Texas, standard succession and its bearing on Upper Ordovician biogeography: Lethaia, v. 28, p. 115-128.

Hardage, B.A., Pendleton, V.M., and Asquith, G.B., 1999, 3-D Seismic Interpretation of Deep, Complex Structures in the Delaware Basin, West Texas: Bureau of Economic Geology Geologic Circular 99-1, 42 p.

Herald, F.A., ed., 1957, Occurrence of oil and gas in West Texas: The University of Texas at Austin, Bureau of Economic Geology Publication 5716, 442 p.

Howe, H.J., 1959, Montoya group stratigraphy (Ordovician) of Trans-Pecos Texas: American Association of Petroleum Geologists Bulletin, v. 43, no. 10, p. 2285-2332.

Holland, S.M. and Patzkowsky, M. E., 1997, Distal orogenic effects on peripheral bulge sedimentation: Middle and Upper Ordovician of the Nashville Dome: Journal of Sedimentary Research, v. 67, no. 2, p. 250-263.

Holmden, C., Creaser, R.A., Muehlenbachs, K., Leslie, S., and Bergstrom, S.M., 1998, Isotopic evidence for geochemical decoupling between ancient epeiric seas and bordering oceans: Implications for secular curves: Geology, v. 26, no. 6, p. 567-570.

Huffman, G.G., 1959, Pre-Desmonian isopachous and paleogeographic studies in central mid-continent region: American Association of Petroleum Geologists Bulletin, v. 43, no. 11, p. 2541-2574.

Jones, T.S., 1953, Stratigraphy of the Permian Basin of West Texas: West Texas Geological Society Publication 53-29, 63 p.

Klein, C., 1993, Manual of Mineralogy (after James D. Dana), 21st Edition: Wiley, New York, 681 p.

Kottlowski, F.E., 1970, Influence of the Pedernal uplift on sedimentation, *in* Basins of the Southwest, v.1, Symposium of the 10th Annual Meeting Southwest Section of American Association of Petroleum Geologists, February 7-9, 1968, Wichita Falls, Texas: West Texas Geological Society Publication 70-68, p.21-40.

Kottlowski, F.E., Flower, R.H., Thompson, M.L., and Foster, R.W., 1956, Stratigraphic studies of the San Andres Mountains, New Mexico: New Mexico Bureau of Mines and Mineral Resources Memoir 1, p. 7-27.

Measures, E.A., 1985, Carbonate facies of the Montoya Group – Description of a shoaling-upward ramp, Part I: West Texas Geological Society Bulletin, v. 25, no. 2, p. 4-8.

Measures, E.A. 1985, Carbonate facies of the Montoya Group – Description of a shoaling-upward ramp, Part II: West Texas Geological Society Bulletin, v. 25, no. 3, p. 4-8.

Measures, E.A., 1984, Petrography, depositional environments, and dolomitization of the Ordovician Montoya Group, New Mexico and Texas [Master's thesis]: Sul Ross State University, Alpine, Texas, 198 p.

Mears, C.E. and Dufurrena, C.K., 1984, Pre-Leonardian geology of Midland Farms field area, Andrews County, Texas, Transactions Southwest Section American Association of Petroleum Geologists, 1984 Convention, Midland, Texas: West Texas Geological Society Publication 84-78, p. 111-123.

McBride, E.F., 1970, Stratigraphy and origin of Maravillas Formation (Upper Ordovician), West Texas: American Association of Petroleum Geologists Bulletin, v. 54, no. 9, p. 1719-1745.

Mitchell, J.S., 2003, Facies analysis and chert geochemistry of Late Ordovician cherty carbonates on the southern margin of Laurentia [M.S. thesis]: Washington State University, 71 p.

Nielsen, A.T., 2004, Ordovician Sea level Changes: A Baltoscandian Perspective, in Webby, B.D., Paris, F., Droser, M.L., and Percival, I.G. (eds.), The Great Ordovician Biodiversification Event: Columbia University Press, p. 84-93.

Northcutt, R.A. and Johnson, K.S., 1997, Major Simpson and Viola oil and gas reservoirs in Oklahoma, *in* Johnson, K.S. (ed.), Simpson and Viola Groups in the southern Midcontinent, 1994 Symposium: Oklahoma Geological Survey Circular 99, p. 48-57.

Pope, M.C., 2004a, Cherty carbonate facies of the Montoya Group, southern New Mexico and western Texas and its regional correlatives: A record of Late Ordovician paleoceanography on southern Laurentia: Palaeogeography, Palaeoclimatology, Palaeoecology 210, p. 367– 384.

Pope, M.C., 2004b, Upper Ordovician and Lower to Middle Silurian Miogeoclinal Rocks, *in* Mack, G.H. and Giles, K.A. (eds.), The Geology of New Mexico, A Geologic History: New Mexico Geological Society Special Publication 11, p. 45-58.

Pope, M.C., 2004c, Upper Ordovician Montoya Group outcrops in the Cooks Range of New Mexico and Southern Franklin Mountains of Texas: Permian Basin Section-SEPM Fieldtrip to accompany 2004 Southwest Section American Association of Petroleum Geologist Annual Meeting in El Paso, Texas, 22 p.

Pope, M.C. and Steffen, J.B., 2003, Widespread, prolonged late Middle to Late Ordovician upwelling in North America: A proxy record of glaciation?: Geology, v. 31; no. 1; p. 63–66.

Pope, M.C., 2002a, Early Ordovician El Paso Formation and Late Ordovician Montoya Formation, Rhodes Canyon, *in* Lueth, V.; Giles, K.A.; Lucas, S.G.; Kues, B.S.; Myers, R.G.; Ulmer-Scholle, D. (eds.), Geology of White Sands: New Mexico Geological Society Guidebook 53, p. 35-39.

Pope, M.C., 2002b, Cherty facies of the Late Ordovician Montoya Group, southern New Mexico and western Texas: Implication of Laurentia oceanography and duration of Gondwana glaciation, *in* Lueth, V.; Giles, K.A.; Lucas, S.G.; Kues, B.S.; Myers, R.G.; Ulmer-Scholle, D. (eds.), Geology of White Sands: New Mexico Geological Society Guidebook 53, p. 159-165.

Pray, L.C., 1958, Stratigraphic section, Montoya Group and Fusselman Formation, Franklin Mountains, Texas: West Texas Geological Society Guidebook, p. 30-42.

Pratt, W.P. and Jones, W.R., 1961, Montoya dolomite and Fusselman dolomite in Silver City region, New Mexico: American Association of Petroleum Geologists Bulletin, v. 45, no. 4, p. 484-500.

Read, J.F., 1995, Part 1. Overview of carbonate platform sequences, cycle stratigraphy and reservoirs in greenhouse and ice-house worlds, *in* Read, J.F., Kerans, C., and Weber, L.J. (eds.), Milankovitch sea level changes, cycles and reservoirs on carbonate platforms in greenhouse and ice-house worlds: SEPM Short Course Notes 35, p. 1-102.

Ross, R.J. and Ross, C.A, 1992, Ordovician sea-level fluctuations, *in* Webby, B.D. and Laurie, J.R. (eds.), Global Perspectives on Ordovician Geology, Proceedings of the Sixth Annual International Symposium on the Ordovician, University of Sydney, Australia, July 15-19, 1991, p. 327-335.

Ross, R.J., 1976, Ordovician sedimentation in the western United States, *in* Bassett, M.G. (ed.), The Ordovician System: Proceedings of a Palaeontological Association Symposium, University of Wales Press and National Museum of Wales, p. 73-105.

Saltzman, M.R. and Young, S.A., 2005, Long-lived glaciation in the Late Ordovician? Isotopic and sequence-stratigraphic evidence from western Laurentia: Geology, v. 33, no. 2, p. 109-112.

Sloss, L.L., 1988, Tectonic evolution of the craton in Phanerozoic time, *in* Sloss, L. L. (ed.), Sedimentary cover—North American Craton: The Geology of North America, Geological Society of America, Boulder, Colorado, v. D-2, p. 25–51.

Sloss, L.L., 1963, Sequences in the cratonic interior of North America: Geological Society of America Bulletin, v.74, no. 2, p. 93-113.

Sternbach, C.A., 1984, Deep-burial diagenesis and Dolomitization in the Hunton Group carbonate rocks (Upper Ordovician to Lower Devonian) in the Anadarko Basin of Oklahoma and Texas [Ph.D. dissertation]: Rensselaer Polytechnic Institute, 144 p.

Sweet, W.C., 1979, Late Ordovician conodonts and biostratigraphy of the western mid-continent province: Geology Studies, v. 26, Brigham Young University, p. 45-85.

Swift, D.B., Reeves, J.J., Erdlac, R.J., Jr., Barrow, K.T., and Schoellhopf, A.P., 1994, Exploration potential of the Diablo Platform-Delaware Basin margin, Trans-Pecos Texas, in Ahlen, J., Peterson, J., and Bowsher, A.L. (eds.), Geologic activities in the 90s, Southwest Section of AAPG 1994, Ruidoso, New Mexico: New Mexico Bureau of Mines and Mineral Resources Bulletin, no. 150, p. 53-69.

Texas Railroad Commission, 2008, Field production data accessed from online database at http://webapps.rrc.state.tx.us/PDQ/home.do on February 26, 2008.

Texas Water Development Board, 1972, A survey of the subsurface saline water of Texas, v. 1, 113 p.

Thomas, D.M. and Liu, H., 2003, Upper Ordovician Montoya sequence stratigraphy and chert porosity in southeastern Delaware basin, *in* Hunt, T. J., and Lufholm, P. H. (eds.), The Permian Basin: Back to Basics: West Texas Geological Society Publication 03-112, p. 37–43.

Webby, B.D., Cooper, R.A., Bergstrom, S.M., and Paris, F. 2004, Stratigraphic framework and time slices, *in* Webby, B.D., Paris, F., Droser, M.L., and Percival, I.G. (eds.), The Great Ordovician Biodiversification Event, Columbia University Press, p. 41-47.

Williams, J.A., 1977, Characterization of oil types in the Permian Basin: text of talk presented to Southwest Section, American Association of Petroleum Geologists, Abilene, Texas, March 7, 1977.

Wright, W.F., 1979, Petroleum geology of the Permian Basin: West Texas Geological Society Publication 79-71, 98 p.

Young, S.A., Saltzman, M.R., and Bergström, S.M., 2005, Upper Ordovician (Mohawkian) carbon isotope (δ13C) stratigraphy in eastern and central North America: Regional expression of a perturbation of the global carbon cycle: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 222, p. 53-76.

FIGURES

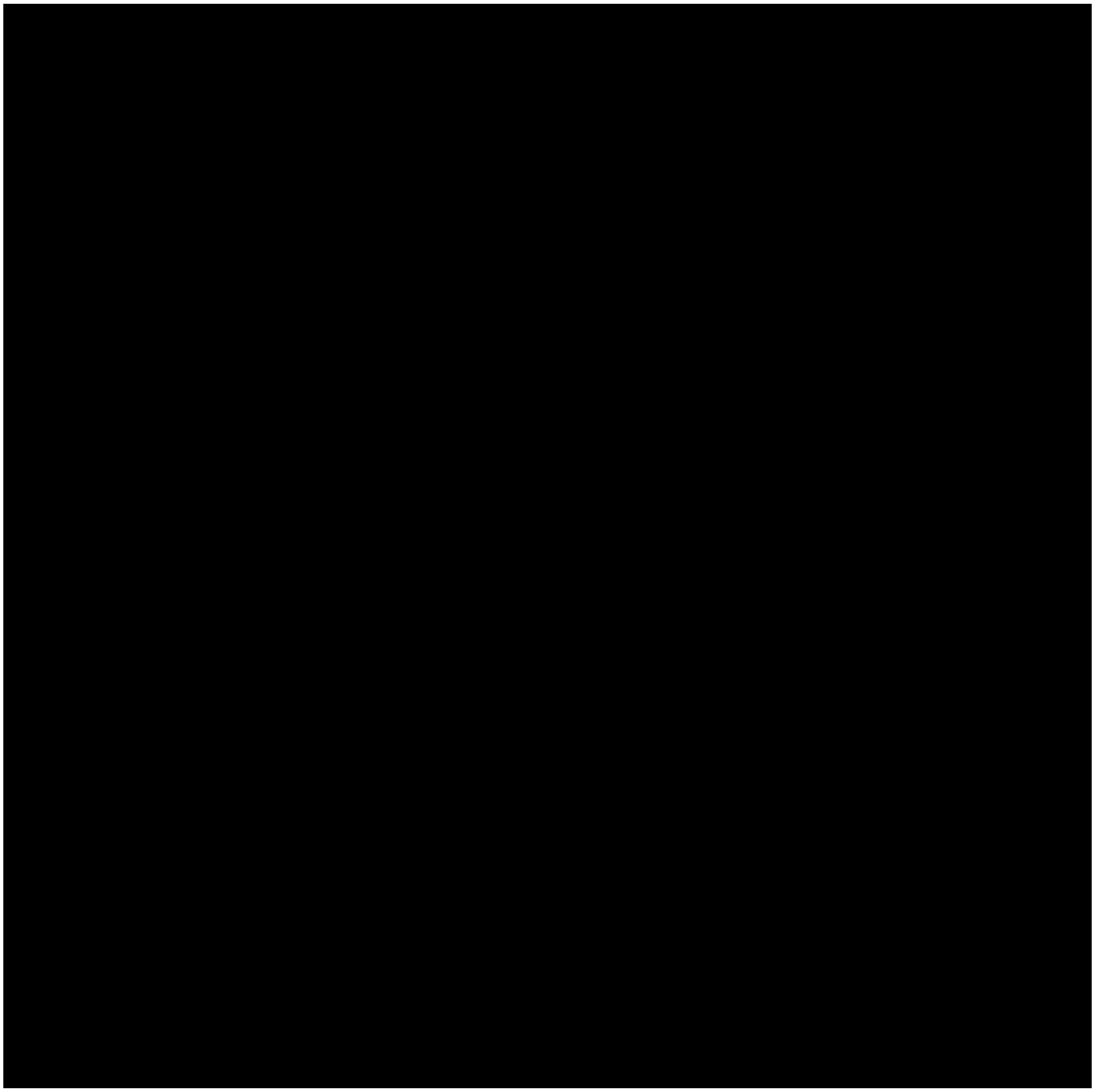

Figure 1. Montoya Group outcrop/subcrop map Outcrop locations after Pope, 2004; subcrop data from published maps in the following regions: Marathon area (Texas Water Development Board, 1972); New Mexico (Frenzel and others, 1988); and Oklahoma (Huffman, 1959, Chenoweth, 1966, and Adler and others, 1971). Line of section shows outcrop transect used to develop the model in Figure 5.

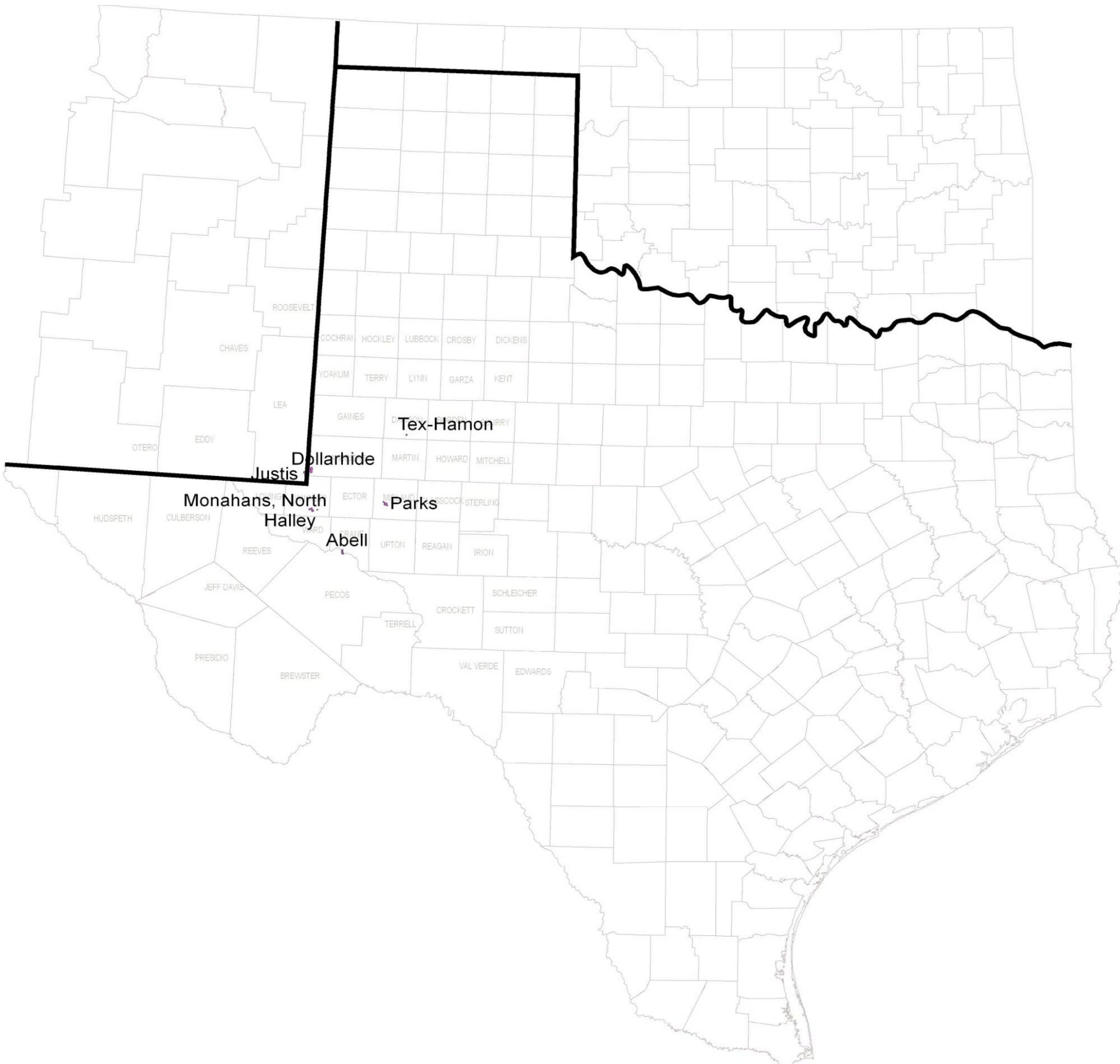

Figure 2. Location map showing Montoya fields with production greater than 1 MMbbl, present subcrop, and outcrops. Producing fields were modified from Dutton and others, 2005; Montoya distribution and regional features were modified from the Texas Water Development Board, 1972; Frenzel and others, 1988; and Northcutt and Johnson, 1997; outcrops were modified from Pope, 2004a.

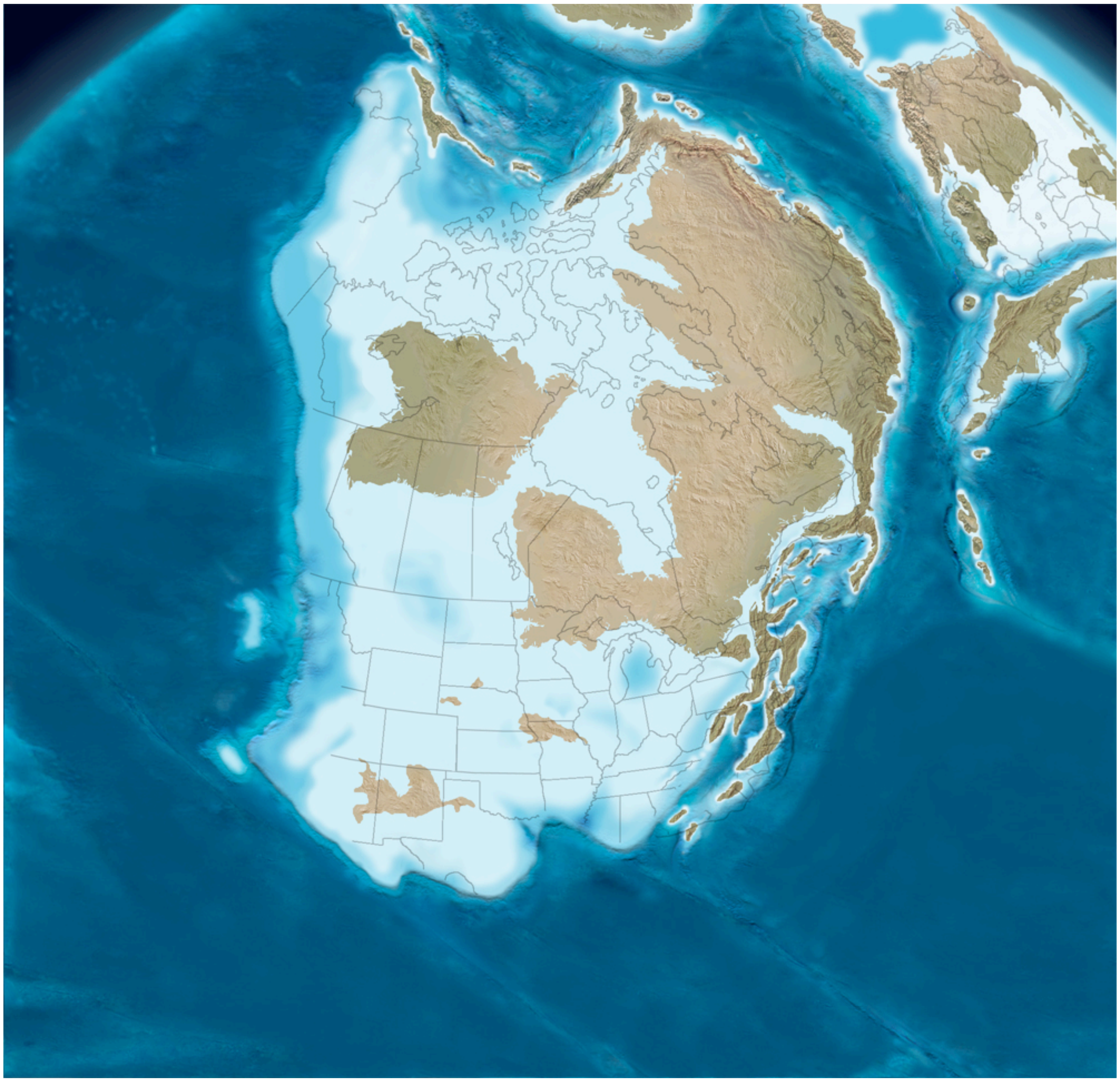

Figure 3. Global plate reconstruction/paleotopography for the Late Ordovician (Blakey, 2004).

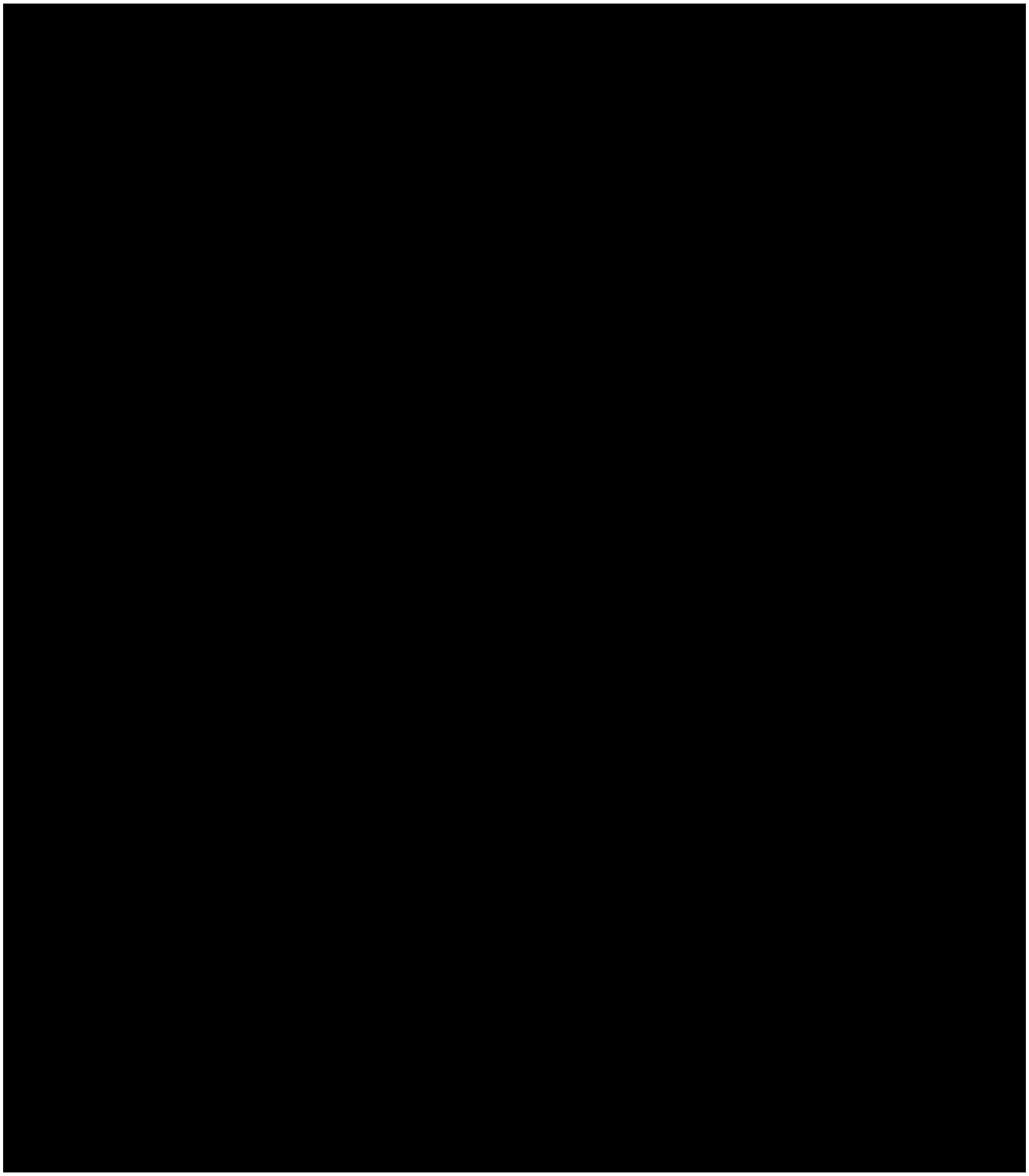

Figure 4. Texas, New Mexico and Oklahoma paleogeography after Ross, 1976.

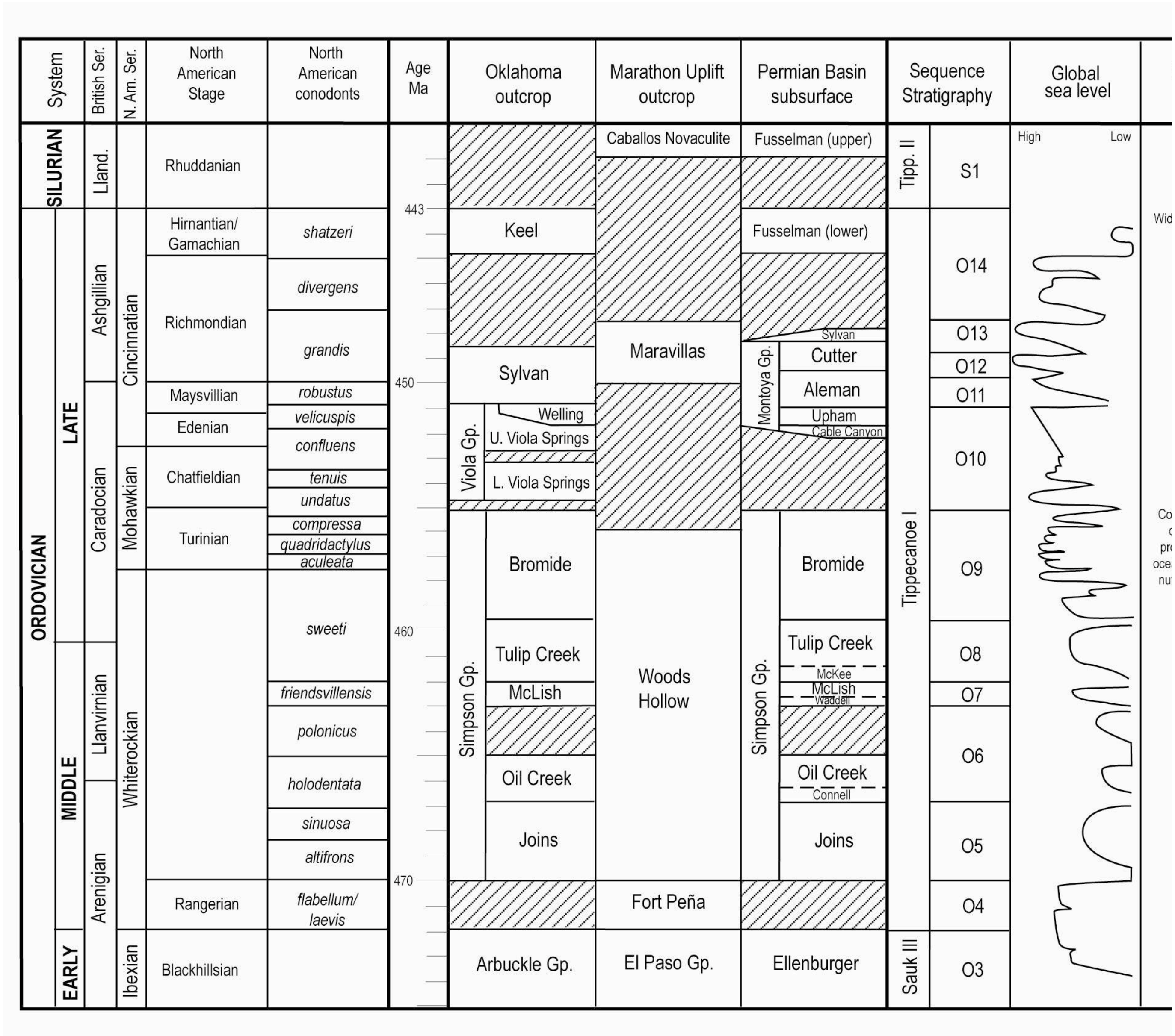

Figure 5. Stratigraphic column. In addition to our own interpretations, sources of information include: Webby and others, 2004 (North American stages, conodonts and time scale); Young and others, 2005 and Derby and others, 1991 (Oklahoma outcrop); Goldman and others, 1995; Bergstrom and others, 1986 (Marathon Uplift outcrop); Pope, 2004 and Sweet, 1979 (Permian Basin subsurface); Sloss, 1988 and 1963 (sequence stratigraphic megasequences); and Ross and Ross, 1992 (global sea level change, note time rescaled to fit biostratigraphy).

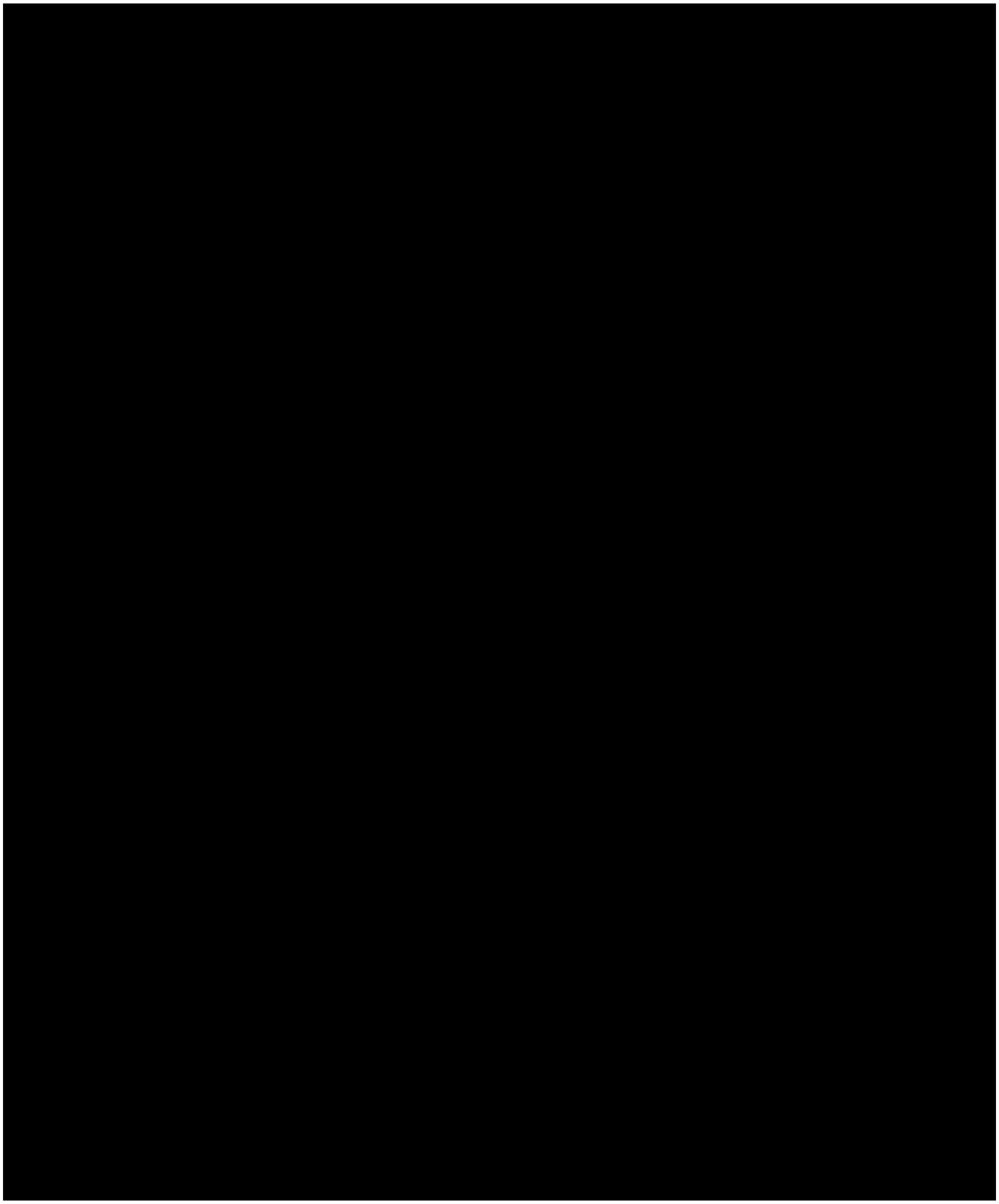

Figure 6. Thickness contours of the Cable Canyon Member in southern New Mexico (after Bruno and Chafetz, 1988) superimposed on paleogeography of the Late Ordovician (after Ross, 1976). Arrows denote interpreted transport direction of sediment eroding from Precambrian basement exposed in the NW to deposition in the SE.

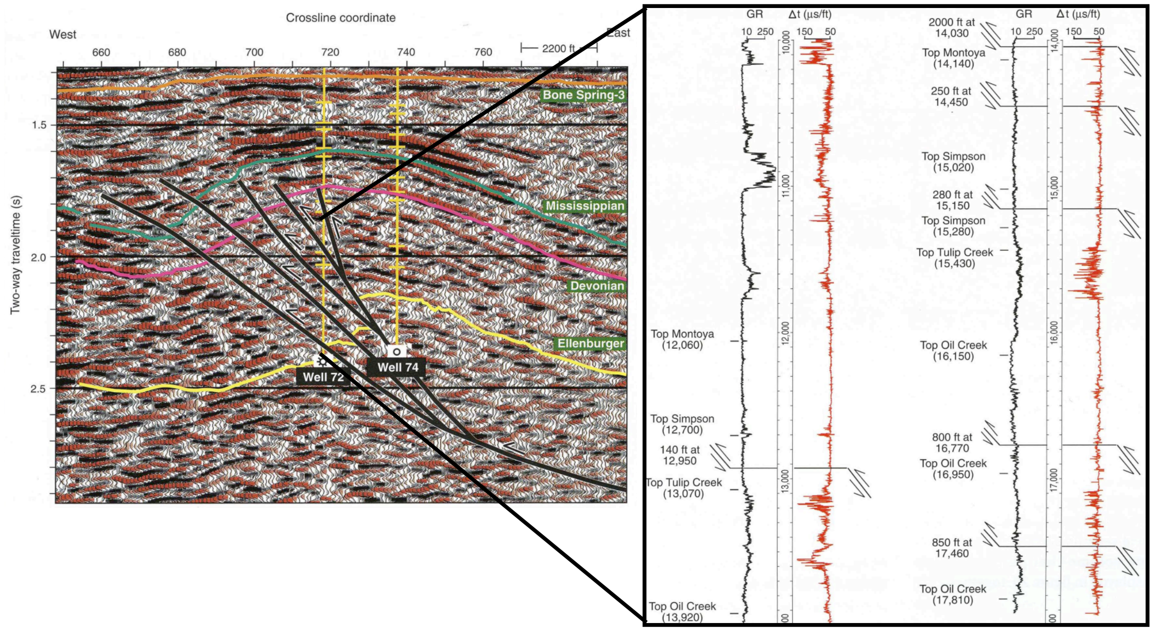

Figure 7. Example of repeated section created through high-angle reverse faulting at Waha Field, Pecos County, Texas (Hardage and others, 1999).

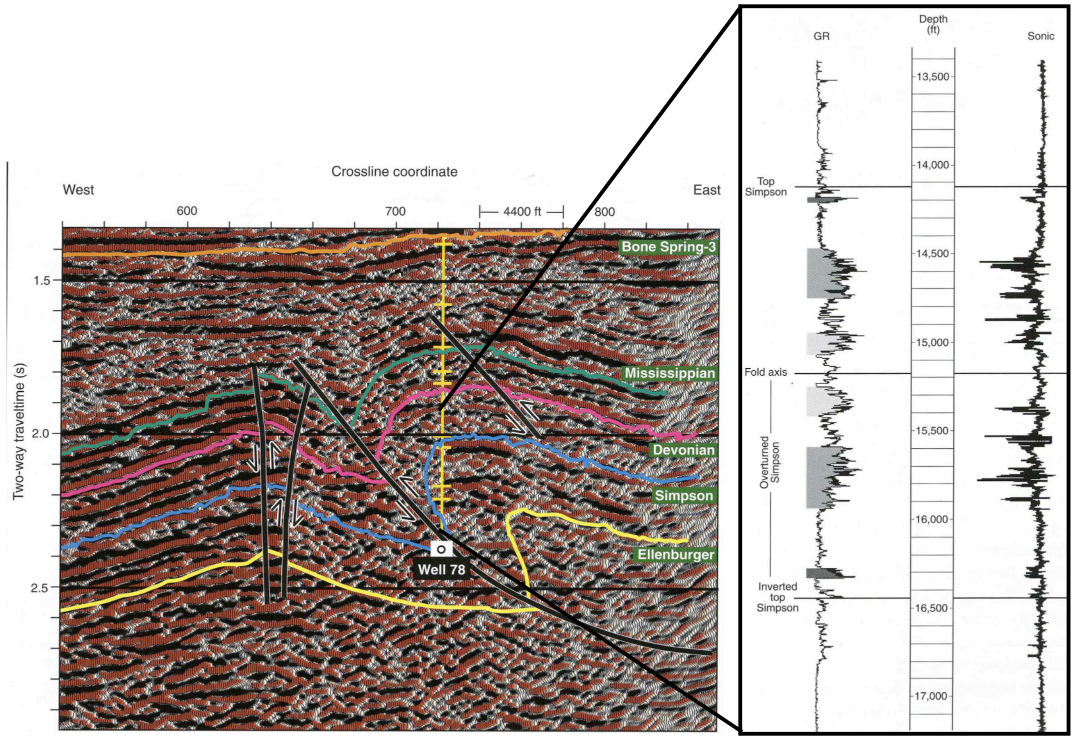

Figure 8. Example of repeated section created through overturned structure at Waha Field, Pecos County, Texas (Hardage and others, 1999).

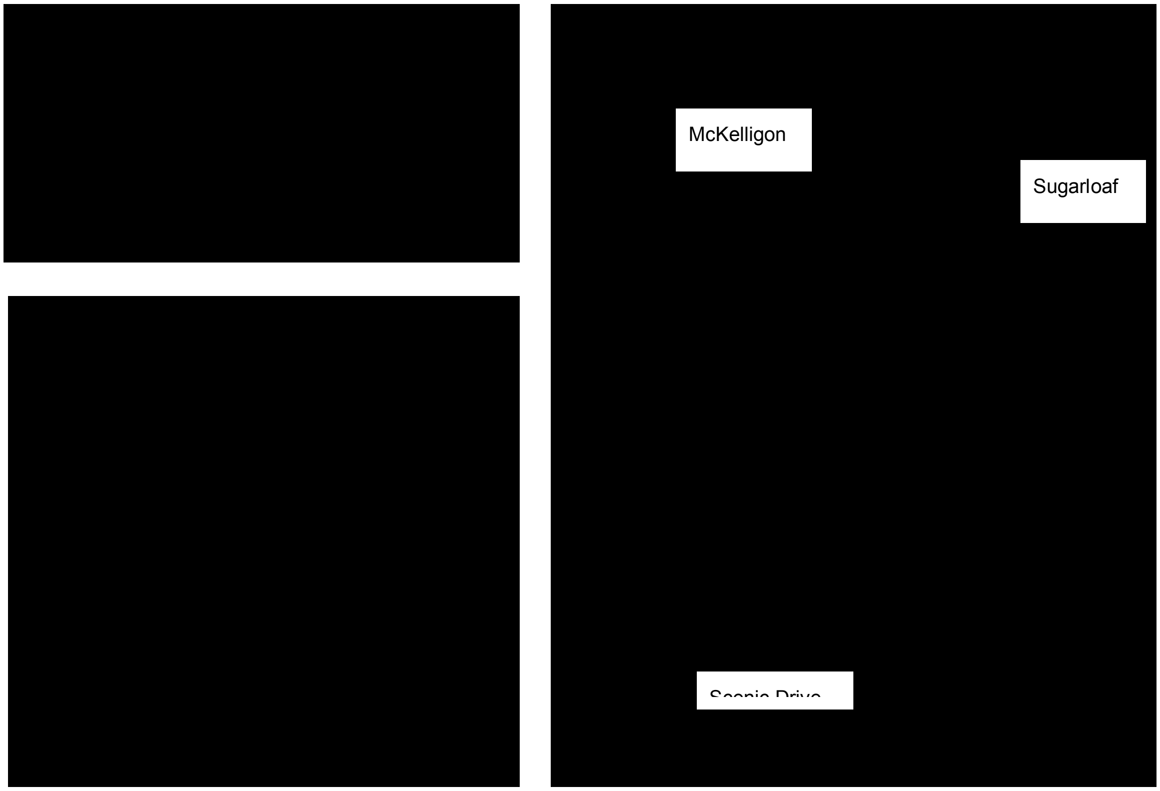

Figure 9. Geologic map of the Franklin Mountains showing Montoya distribution (Om) and field locations (geology from Collins and Raney, 2000).

Figure 10. Third-order sequence stratigraphic model based on Montoya Group outcrops (after Pope, 2004a). Outcrop line of section shown in Figure 1.

Figure 11. Outcrop photo of contact between the more resistant Upham Formation of the Montoya Group overlies the more recessive sandstones of the thin Cable Canyon Member, Franklin Mountains, McKelligon Canyon, El Paso, Texas.

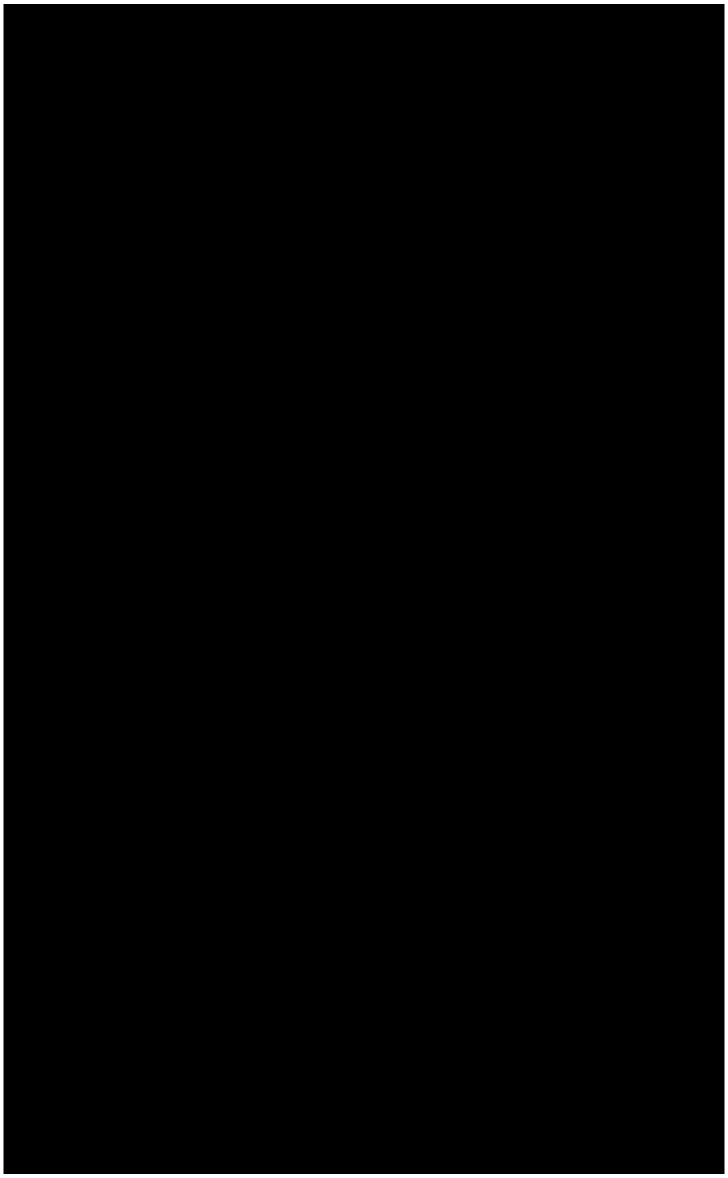

Figure 12. Outcrop photos of large coral in the Upham Formation, Murchison Park stop along Scenic Drive, Franklin Mountains, El Paso, Texas.

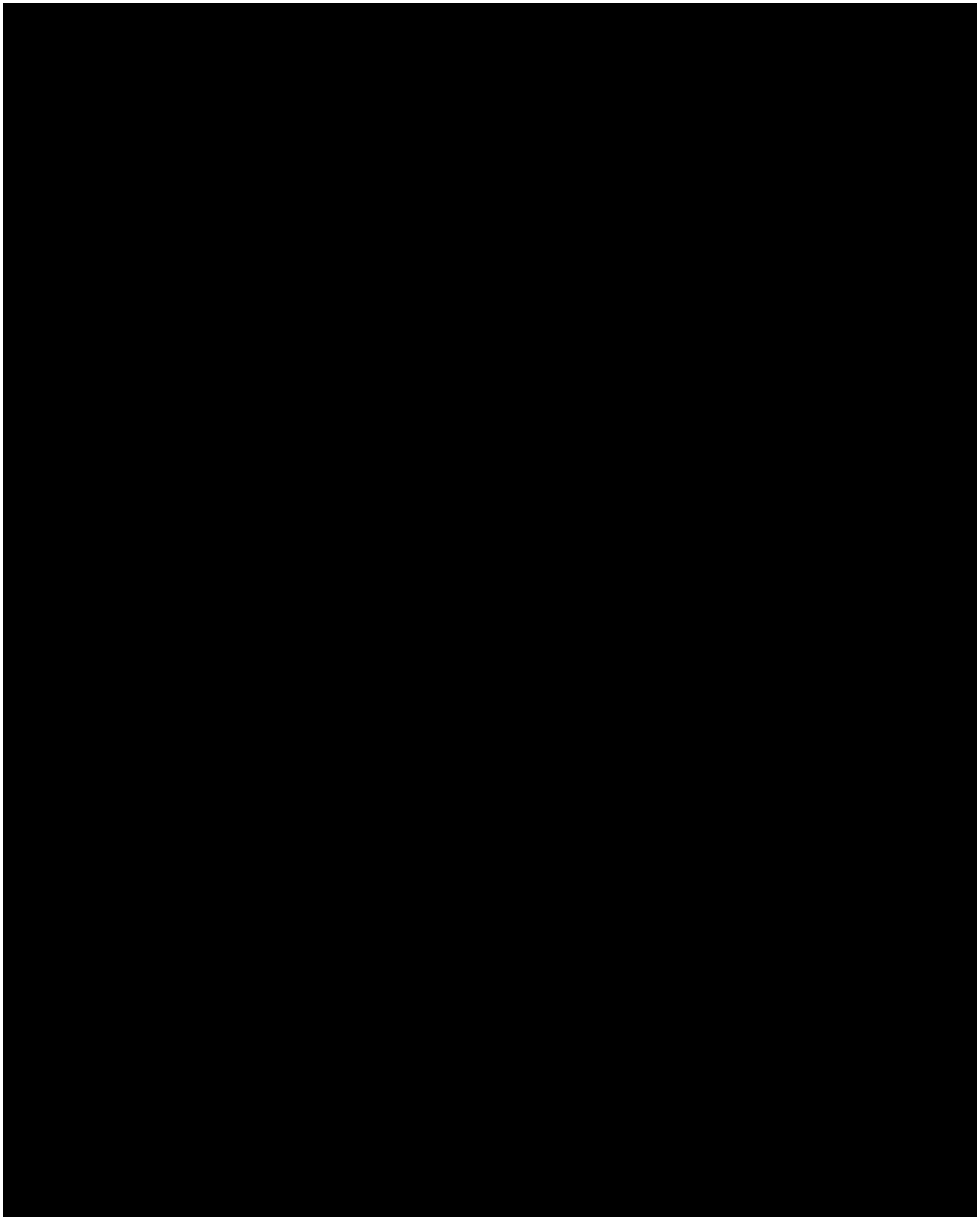

Figure 13. Outcrop photos and corresponding photomicrographs of thin sections in the lower Upham Formation, McKelligon Canyon, Franklin Mountains, El Paso, Texas. Black box in ridge pan denotes area of close-up and thin sections.

A)

[redacted]

B)

[redacted]

Figure 14. A) Second-order transgressive systems tract facies related to Montoya Group Formations (after Pope, 2004a). **B)** Second-order highstand systems tract facies related to Montoya Group Formations (after Pope, 2004a).

Figure 15. Outcrop photos and corresponding photomicrographs of thin sections of the lower Aleman Formation thin-bedded chert, McKelligon Canyon, Franklin Mountains, El Paso, Texas. Thin section photomicrographs show complete dolomitization with no preservation of porosity.

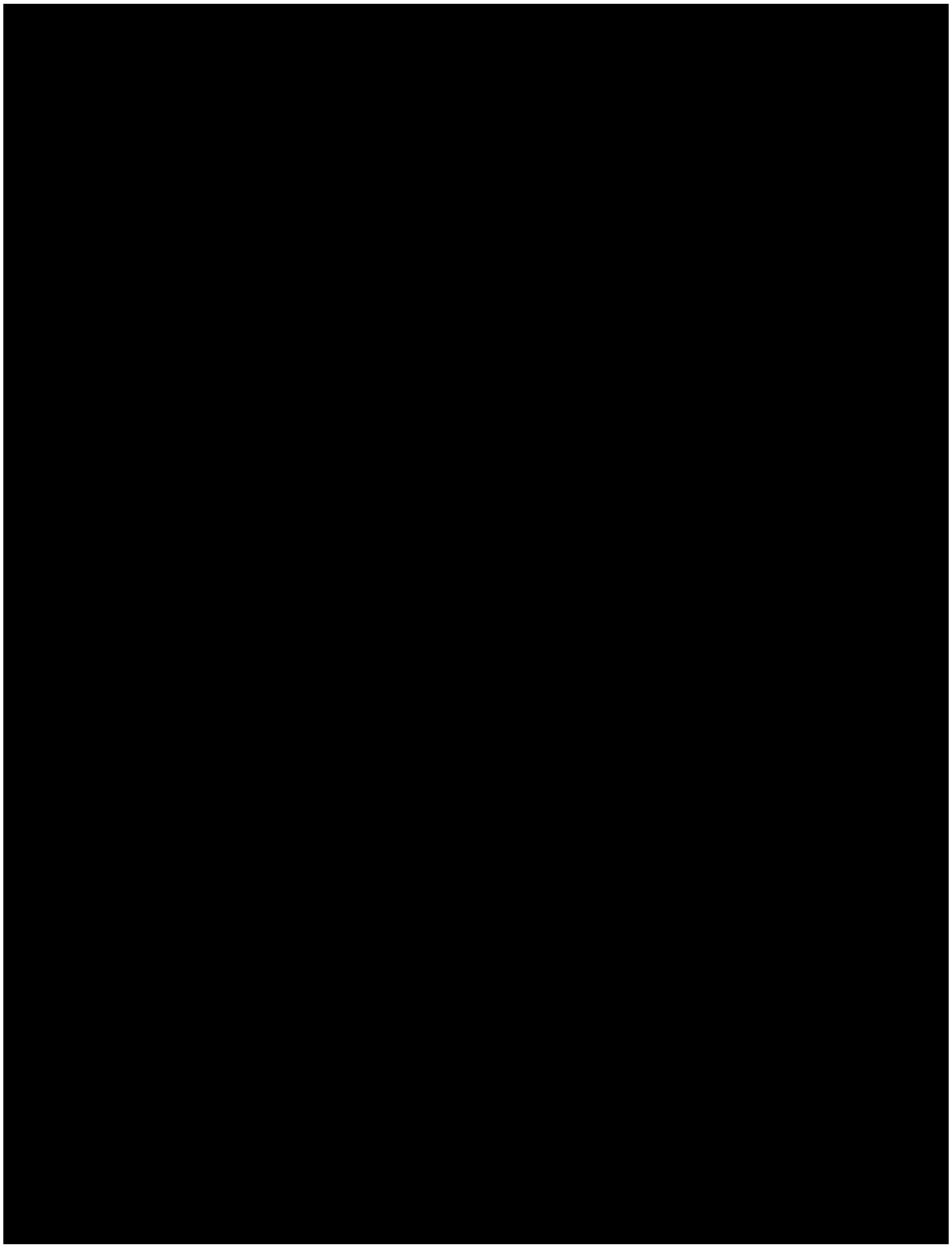

Figure 16. Outcrop photo and corresponding thin section photomicrographs of the Aleman Formation medial bryozoan-rich skeletal packstone, McKelligon Canyon, Franklin Mountains, El Paso, Texas. Note presence of bryozoans both on surface of outcrop and in thin section. Late calcite (stained pink with Alizarin red) partially fills bryozoan molds, but some porosity has been preserved.

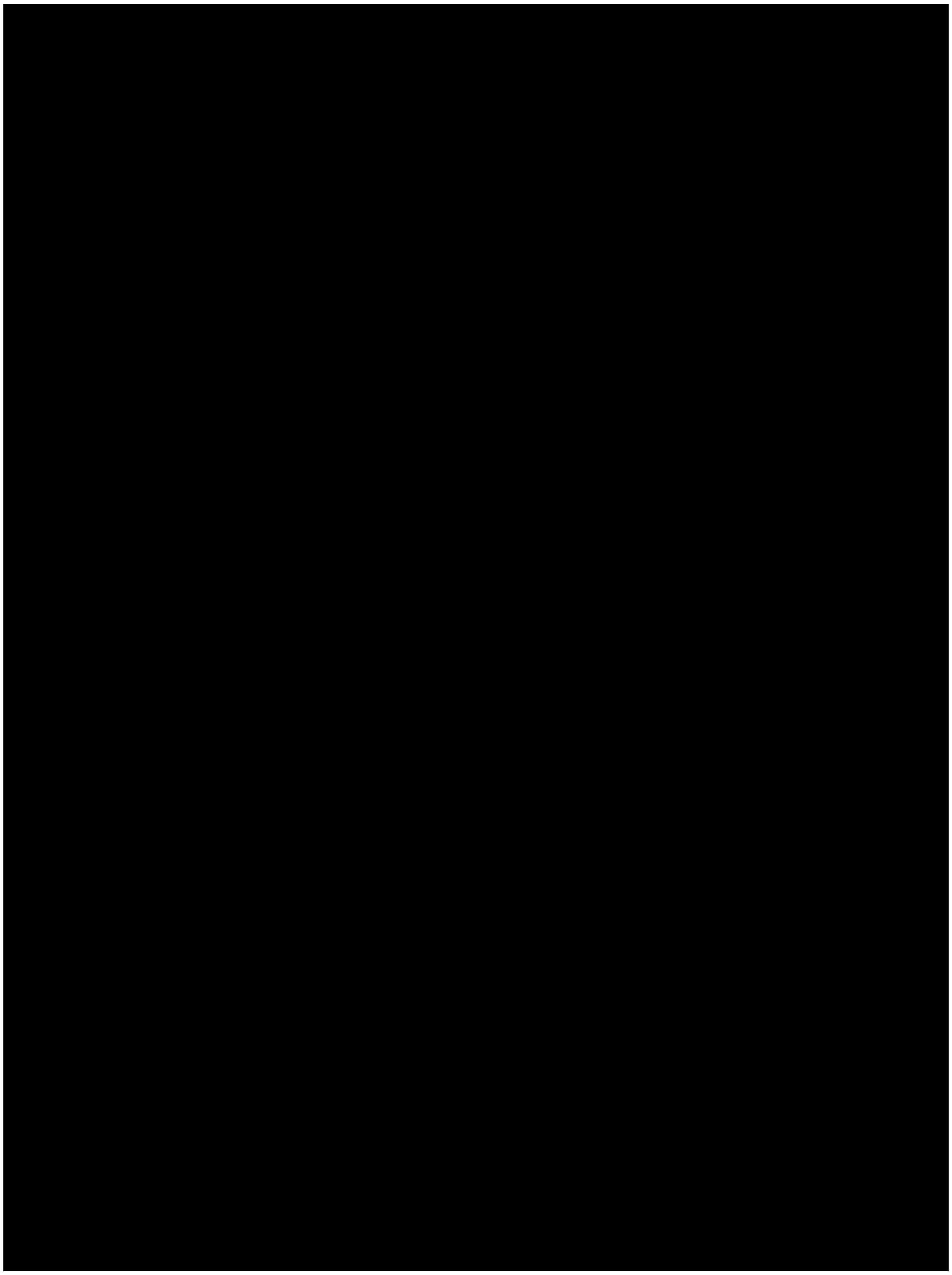

Figure 17. Outcrop photo and corresponding thin section photomicrographs of the upper Aleman Formation chert (chaotic, nodular), McKelligon Canyon, Franklin Mountains, El Paso, Texas. Note sponge spicules, minor microporosity, and dolomite rhombs within chert nodules in photomicrographs.

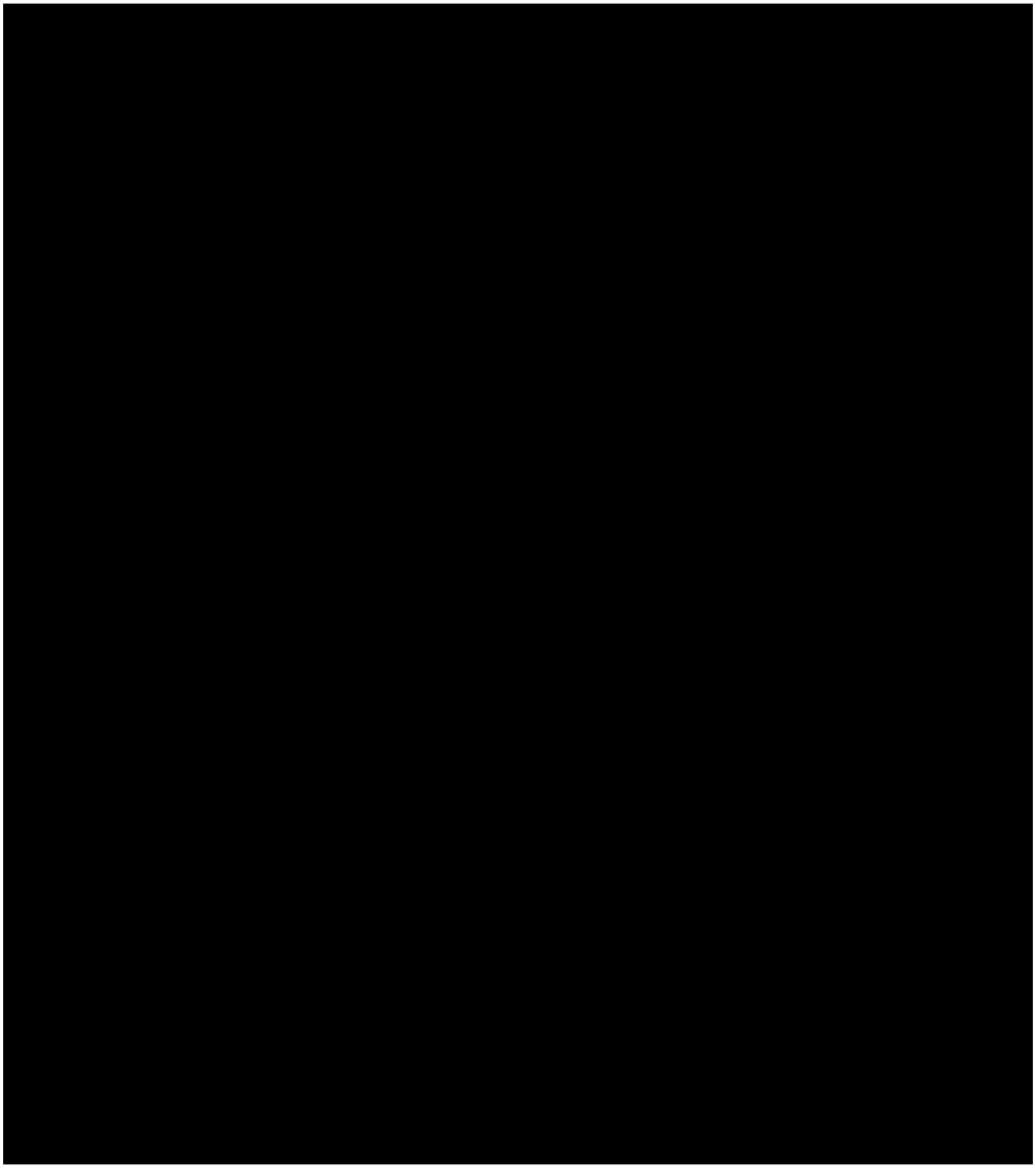

Figure 18. Outcrop photo and corresponding thin section photomicrographs of the lower Cutter Formation, McKelligon Canyon, Franklin Mountains, El Paso, Texas. Photomicrographs show calcite (pink, stained with Alizarin red) filling molds and a fracture.

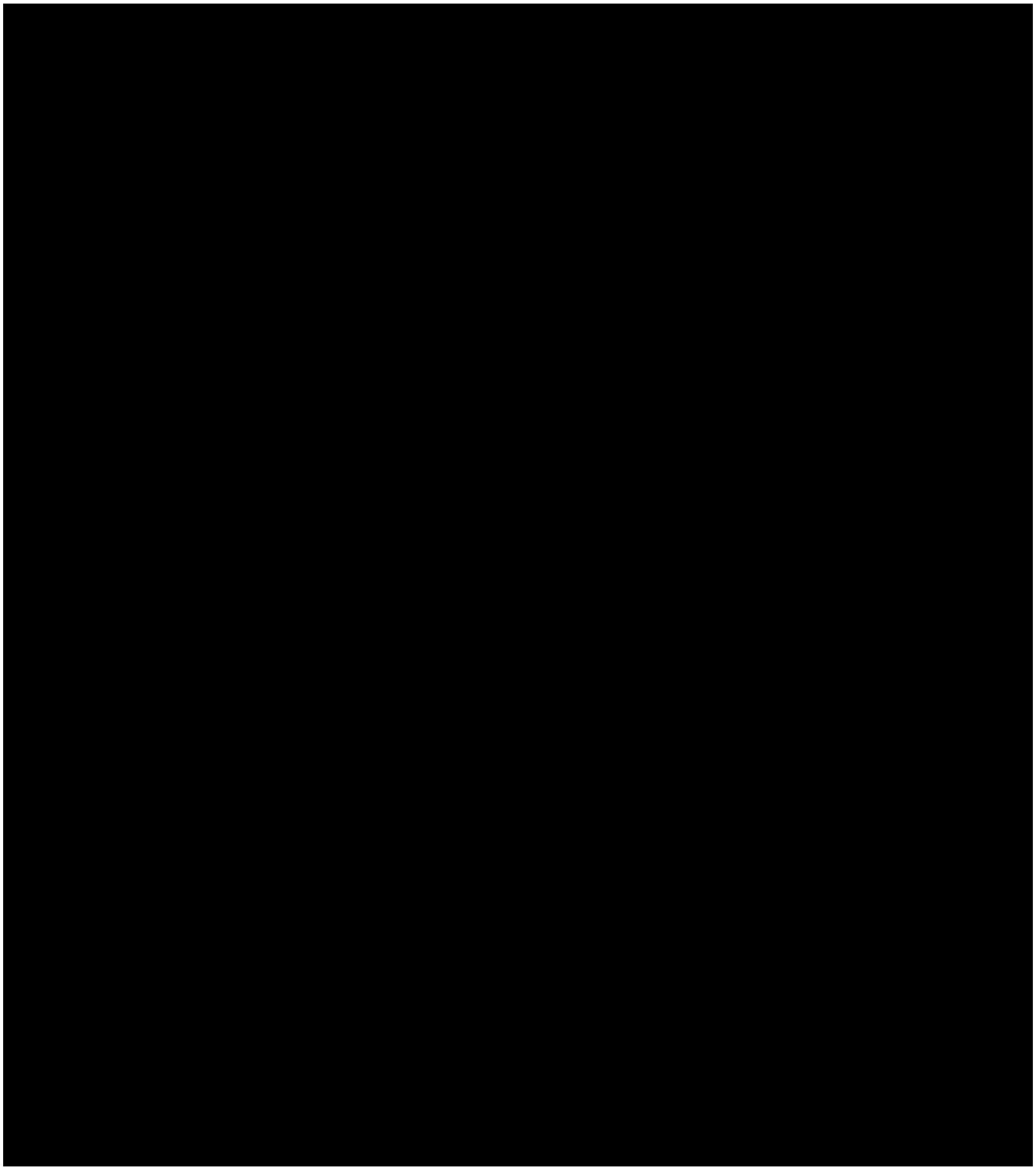

Figure 19. Outcrop photo and corresponding thin section photomicrograph of the lower Cutter Formation, McKelligon Canyon, Franklin Mountains, El Paso, Texas. This location is above Figure 26, but still in the lower Cutter Formation. The outcrop consists of finely laminated mudstone, with laminae more resistant to weathering. The thin section photomicrograph shows laminated fine crystalline dolomite with rare molds.

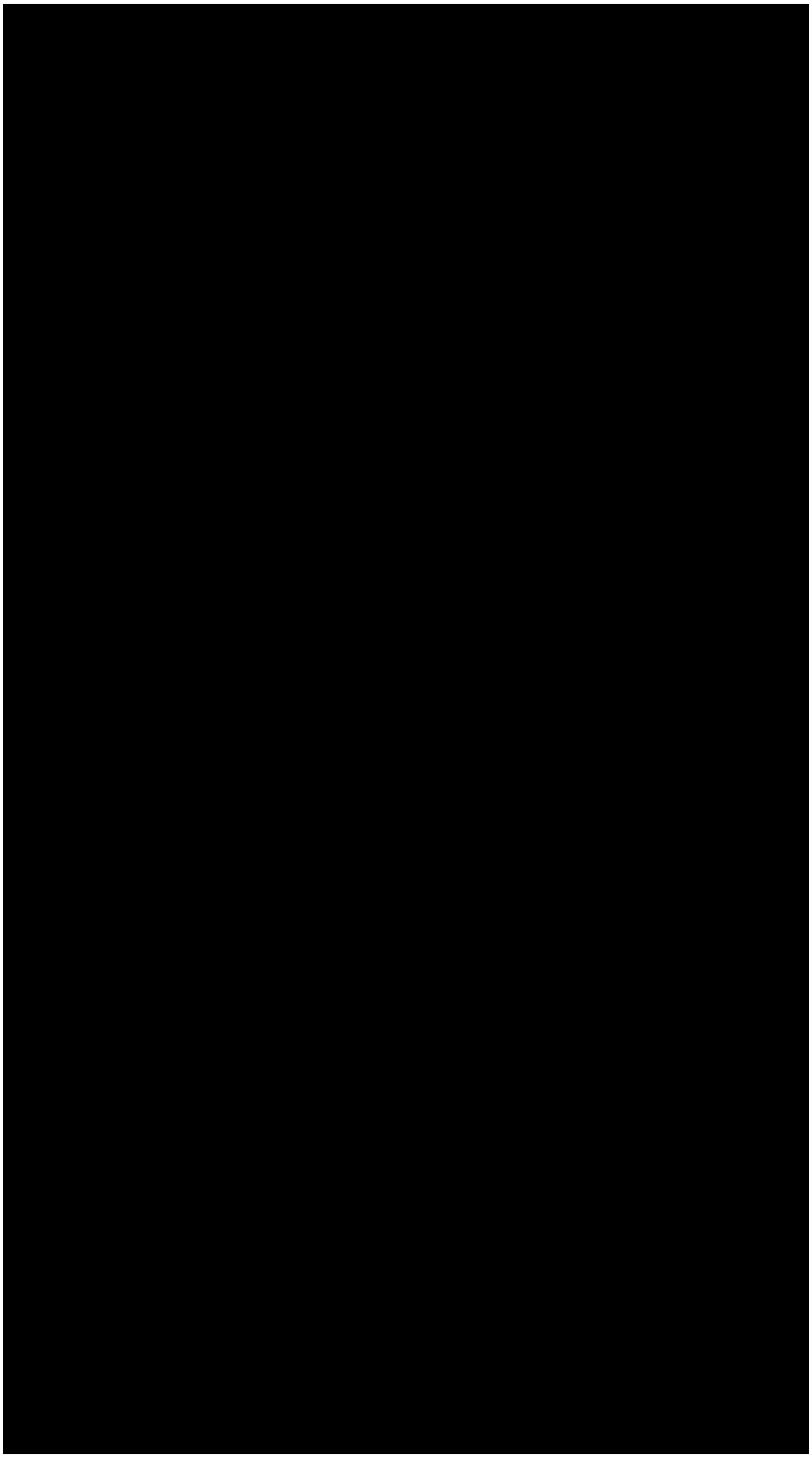

Figure 20. Description of the Montoya cored section in the Dollarhide 25-2-S well.

Figure 21. Map of Dollarhide showing field location, structure (top Fusselman), and core location.

Figure 22. Core photo and thin section photomicrograph of cherty mudstone chert, Dollarhide field 25-2-S, 8541 ft. Photomicrograph shows fine dolomite and fracture partially filled with chert.

Figure 23. Core photo and thin section photomicrograph of crinoid wackestone, Dollarhide field 25-2-S, 8452 ft. Photomicrography shows crinoid fragment (pleochroic in polarized light) and a lack of porosity.

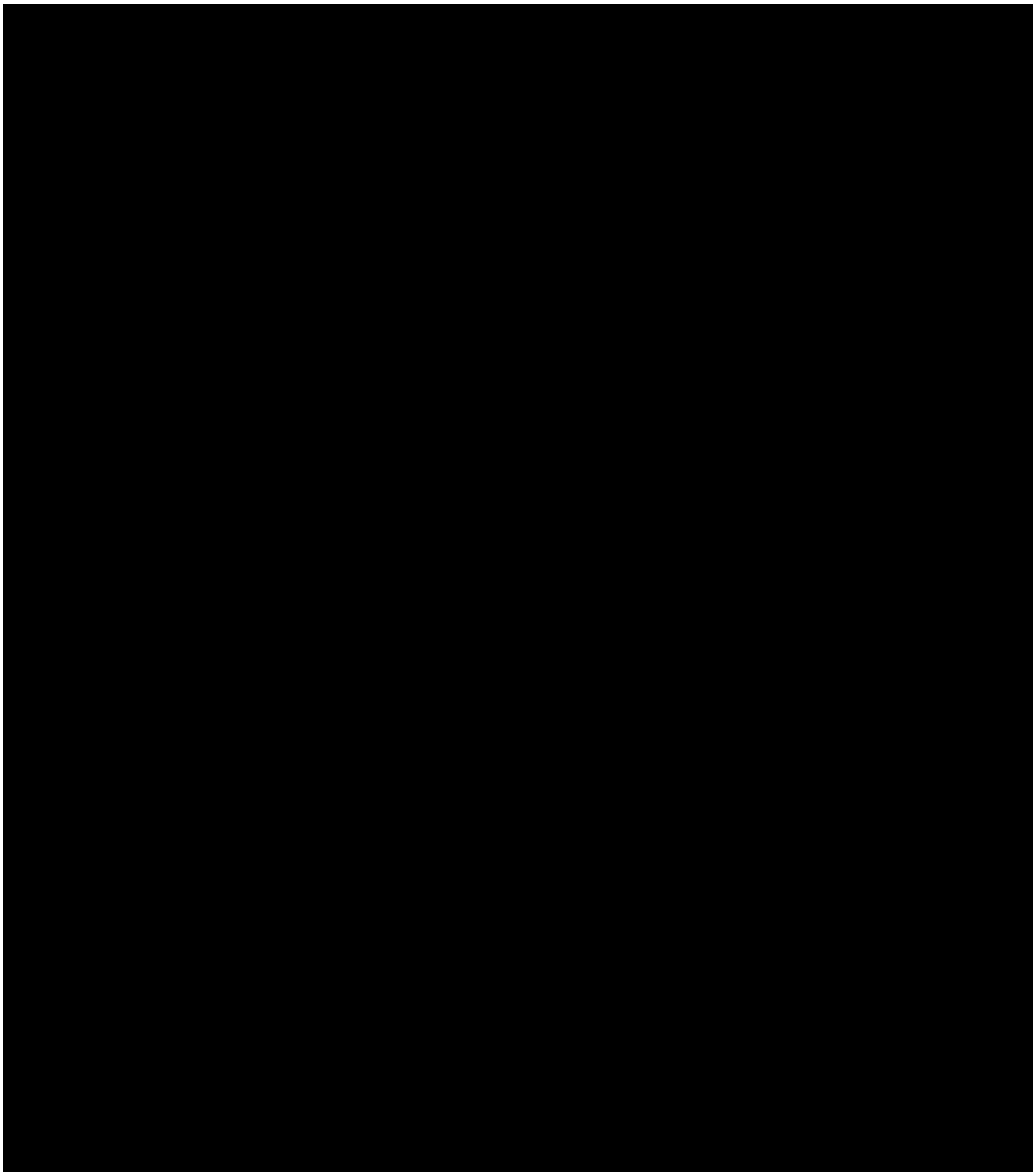

Figure 24. Core photo and thin section photomicrographs of dolopackstone, Dollarhide field 25-2-S, 8487 ft. Coarse dolomite obscures most grains, but peloidal shapes and grain-supported structure are visible. Both photomicrographs show abundant intercrystalline porosity. Whole core porosity = 8.1%, permeability = 40.9 md.

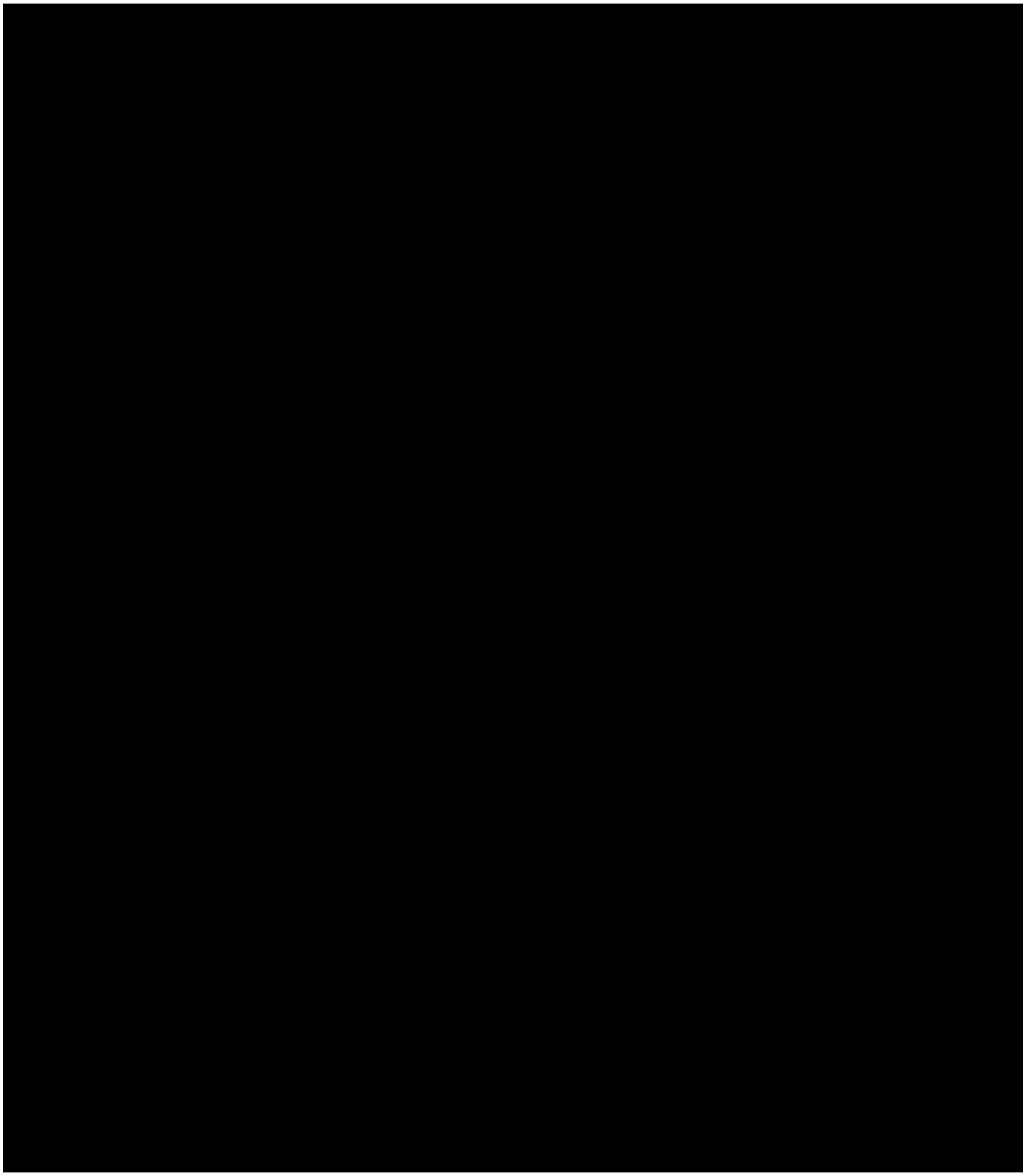

Figure 25. Core photo and thin section photomicrograph of grain-dominate dolopackstone, Dollarhide field 25-2 S, 8527 ft. Grain ghosts are apparent despite coarse crystalline dolomite and abundant intercrystalline porosity. Whole core porosity = 8.4%, permeability = 6.33 md.

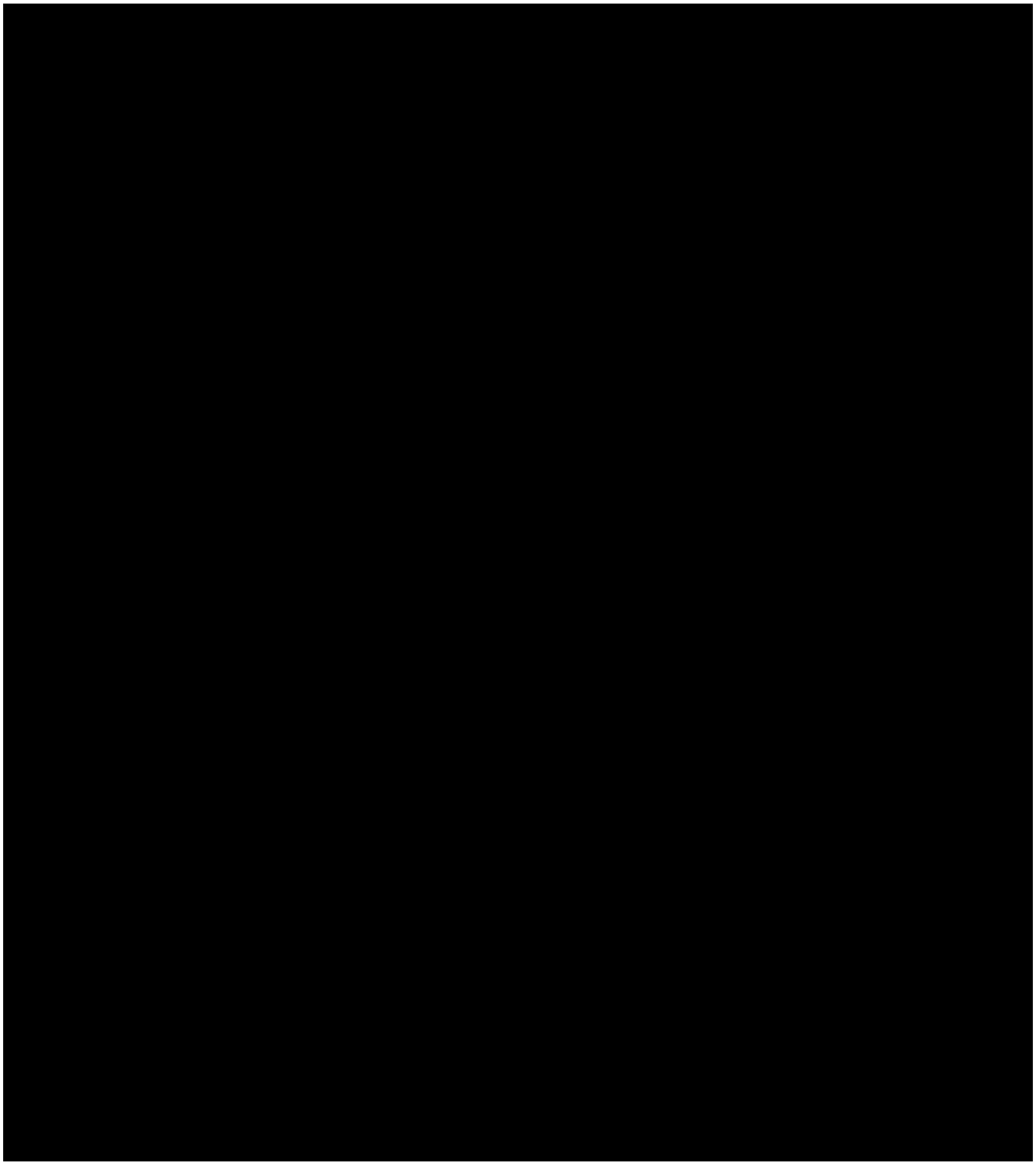

Figure 26. Core photo and thin section photomicrographs of chert mudstone, Dollarhide field 25-2, 8467 ft. Thin section was taken to image chert nodule rim (black box denotes cutting location). The top photomicrograph is in plane light and shows microporosity (light blue) along the lower rim of the chert nodule and hints of sponge spicules within the chert nodule. Sponge spicules are more obvious in polarized light (bottom photomicrograph).

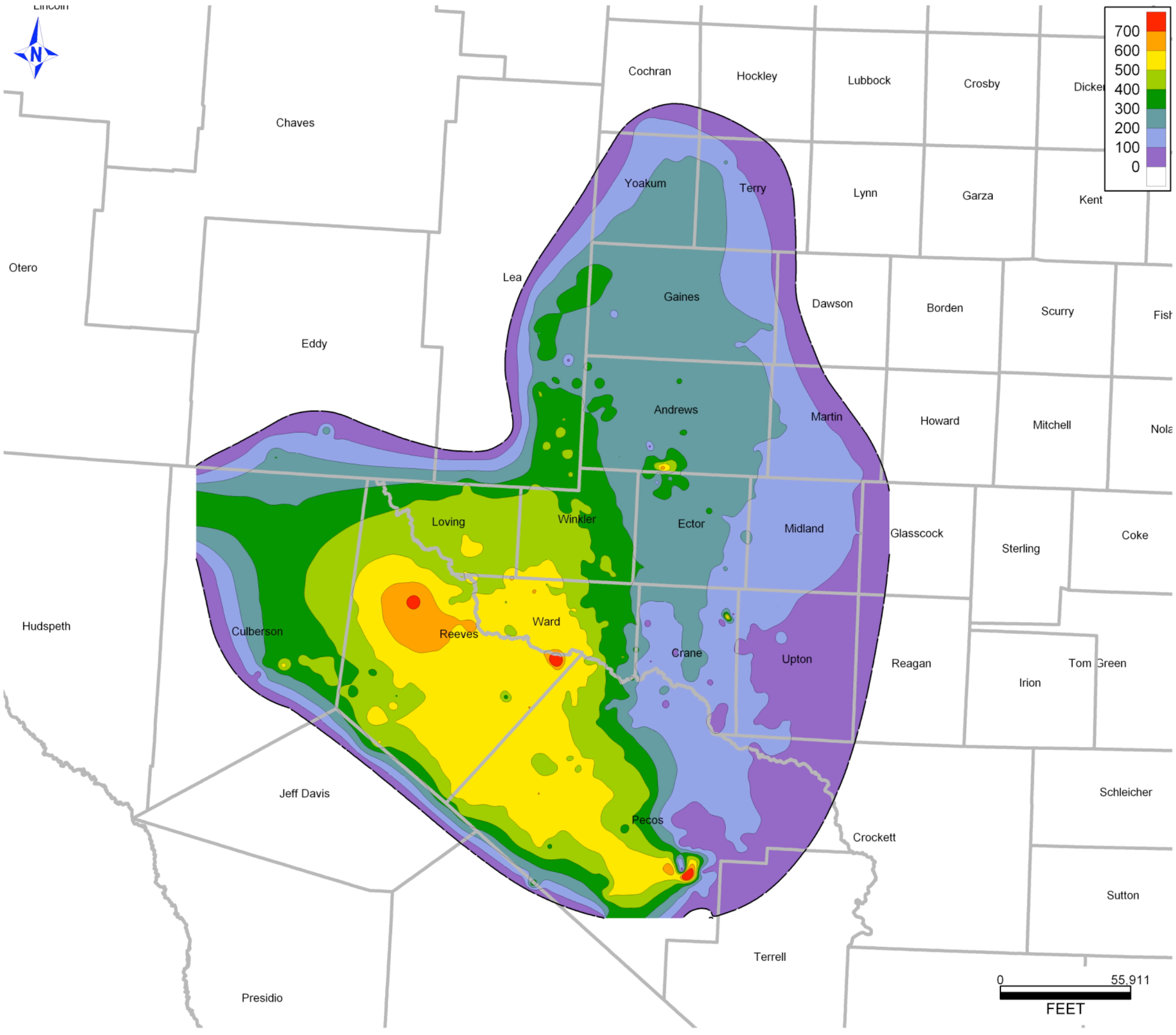

Figure 27. Regional thickness map of Montoya. Data based in part on tops provided by Geological Data Services, Inc. .