THE DEVONIAN SEDIMENTARY RECORD OF MONTANA

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ABSTRACT

More than a century has passed since Peale established the presence of Devonian rocks in Montana, and since then, much has been documented about the distribution, facies character, and depositional origin of these strata. Most attention has been given to the Devonian record in the western part of Montana where rocks are widespread on the surface. However, this record from outcrop is incomplete, since during the Kaskaskia transgression Montana was first inundated in the northeast in what is now the Williston Basin. There the Devonian record goes as far back as the Eifelian with basal conglomerates of the Ashern Formation, representing a transgressive lag deposited on the previously exposed craton. In the western half of Montana widespread Devonian deposition did not occur until the latest Givetian (Maywood Formation) and more prominently during earliest Frasnian. At this time the Kaskaskia transgression closed in from the present-day east and west, depositing the Jefferson Dolomite, the thickest Devonian formation in Montana. This deposition did not occur gradually, but stepwise, resulting in the formation of well-recognized depositional cycles that reflect changes in accommodation on the craton. Accommodation was mainly driven by eustacy during the Eifelian to early Frasnian, but tectonic controls became more significant by the end of the Devonian as the Antler Orogeny influenced the cratonic margin. This transition from a mainly passive margin setting to an active margin resulted in the formation of small sub-basins across present-day Montana. These basins were typically bound by deep-rooted lineaments that were reactivated during the Antler collision. This tectonic inheritance and local deposition is best recorded in the Famennian to Tournaisian strata, namely the Bakken, Exshaw, and Sappington Formations. This review summarizes the Devonian strata in Montana and the diverse observations and interpretations as reported by researchers over more than a century.

INTRODUCTION

"The following note is written simply to place upon record the first positive identification of Devonian strata in the Rocky-Mountain region of Montana." These are the words that A.C. Peale used to describe for the first time the occurrence of Devonian rocks in Montana (Peale, 1885). Nearly a century and a half later, hundreds of peer-reviewed articles, theses, and dissertations, and countless non-peer-reviewed reports, abstracts, and presentations have been published that describe the Devonian strata in Montana at the surface and in the subsurface (fig. 1). The number of Devonian publications in the State vary greatly and are often related to developments in the energy industry. Up until the 1950s, publications about the Devonian in Montana were rare. A first uptick in publications is notable in the 1960s and 1970s, not to a small degree driven by the USGS and the work done by Sandberg and others. A flourish of publications followed in the 1980s, a result of the 1970s oil crisis and the interest in the hydrocarbon potential of several of the Devonian strata in Montana. The number of publications stayed at a relatively high level throughout the next decades, but started to drop from this high nevertheless. The recent uptick in peer-reviewed publications shows that the Devonian in Montana remains an interval of great interest and active geologic research, in no small part because of the "unconventional" oil and gas drilling boom that swept across North America a decade ago, elevating the Williston Basin (fig. 2) in northeastern Montana and adjacent states and Canadian provinces to one of the top three oil-producing basins in the continental United States (Gaswirth and others, 2013; Gaswirth and Marra, 2015).

The "Bakken Boom," named after the main hydrocarbon-bearing latest Devonian Bakken Formation, not only sparked an increase in geologic work in the subsurface in northeastern Montana and adjacent

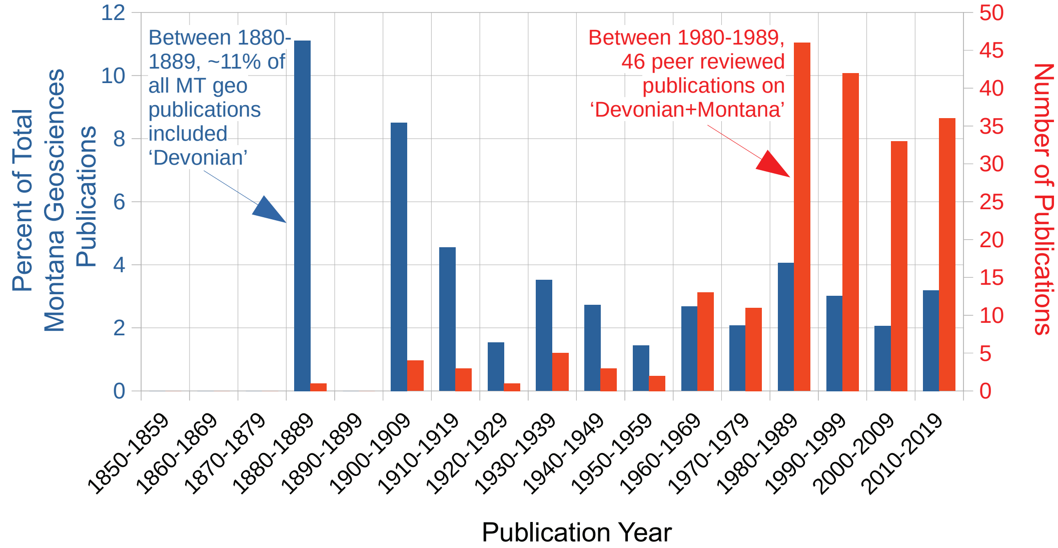


Figure 1. Bar graph showing the number of peer-reviewed publications written about the Devonian in Montana in absolute numbers per decade (red bars) and in percent of all geologic publications published on Montana geology per decade (blue bars). A total of 200 peer-reviewed publications are listed on GeoRef that are concerned with the Devonian in Montana or mention Devonian in Montana (GeoRef search on 'Devonian AND Montana' on July 27th, 2020).

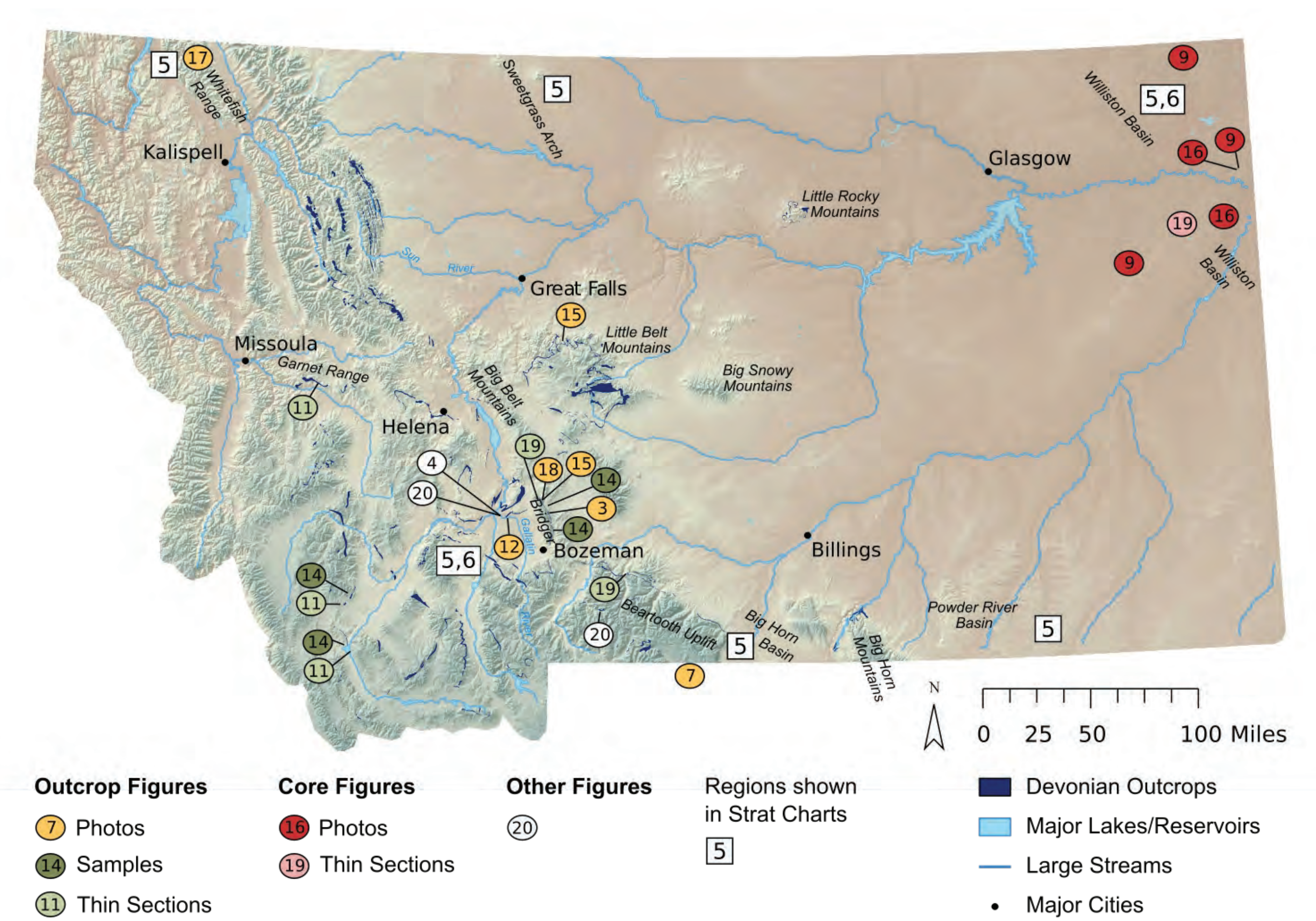


Figure 2. Location map of figures (markers) and locations (italic labels) referred to in this report. Background map is a colored, shaded relief map of Montana, with cooler colors depicting higher elevations.

states and provinces (e.g., Sonnenberg and others, 2017; Hogancamp and Pocknall, 2018; Petty, 2019; Hart and Hofmann, 2020), but it also drove a flurry of studies in the central and western part of the State (e.g., Myrow and others, 2015; Rodriguez and others, 2016; diPasquo and others, 2017; Phelps and others, 2018; Hohman and others, 2019; diPasquo and others, 2019; Browne and others, 2020; Schultz and Hofmann, 2021). There, Devonian strata are well exposed (figs. 2, 3) and are used as an analog to help better understand the stratigraphic architecture and facies distribution that control the production of oil and gas in the subsurface.

In the past couple of decades, sequence stratigraphic analyses and other modern analytical techniques produced a more complex stratigraphic picture of the Devonian in Montana, while also challenging certain assumptions about statewide synchronicity of events, paleogeography, and controls on sedimentation (e.g., Johnson and Sandberg, 1988; Dorobek, 1991; Gantyno, 2010; Grader and others, 2014, 2016; Cole and others, 2015; Myrow and others, 2015; di Pasquo and others, 2017, 2019; Phelps and others, 2018; Hogancamp and Pocknall, 2018; Hohman and others, 2019; Hart and Hofmann, 2020; Browne and others, 2020; Ronemus and others, 2020). This review is an attempt to summarize over a century of excellent Devonian geologic research in Montana with an eye on the large-scale controlling factors, namely eustacy and tectonics.

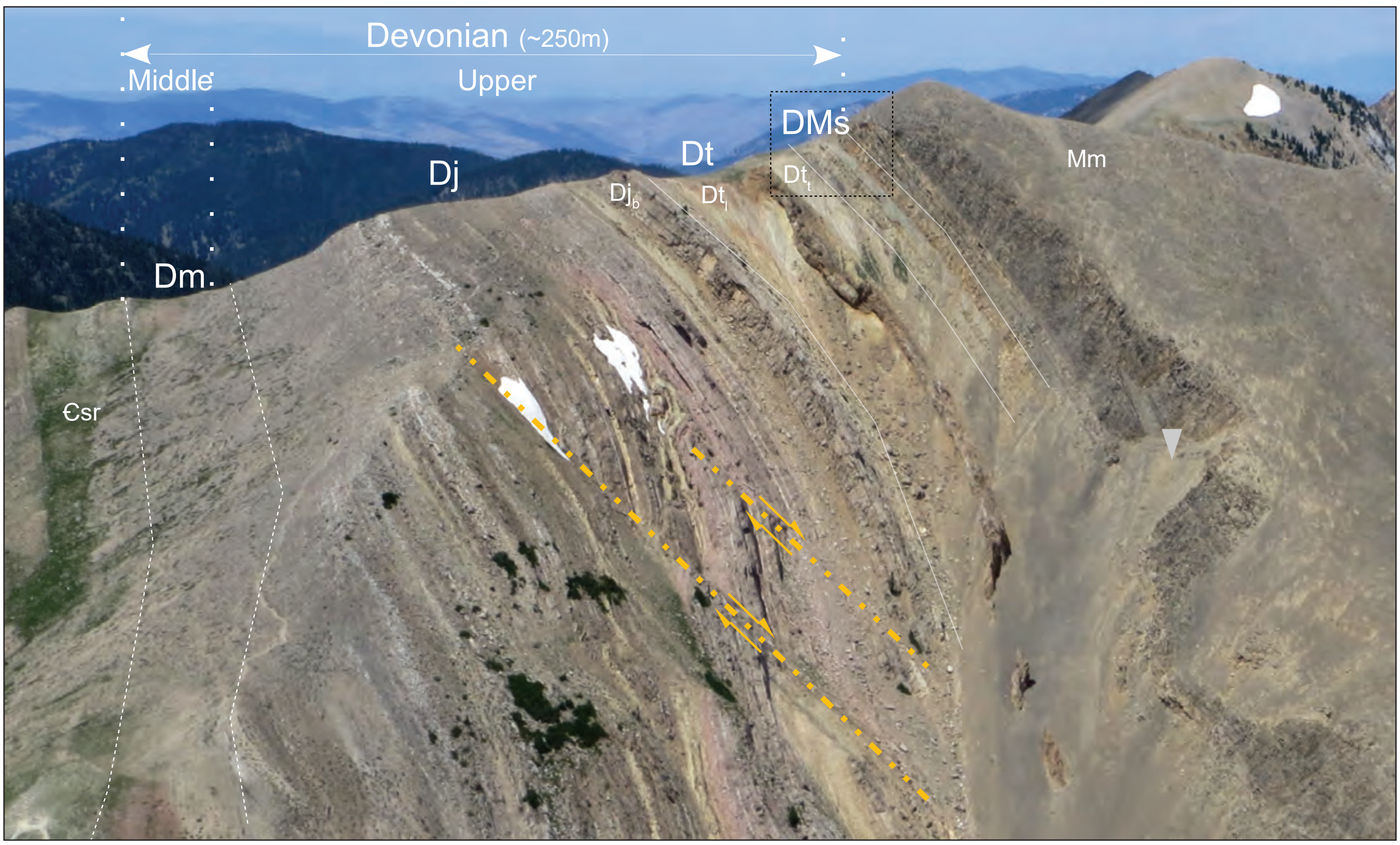


Figure 3. The Devonian strata in the northern Bridger Range in SW Montana as viewed from Sacagawea Peak looking to the northwest. The entire Devonian section is approximately 250 m (820 ft) thick at this location. The oldest Devonian rocks are the Givetian to Frasnian Maywood Formation (Dm) disconformably overlying the Cambrian Snowy Range Formation (Єsr). The Maywood Formation at this location is as much as 27 m thick and primarily composed of reddish orange to pale yellow gray, thin-bedded siltstones and mudstones (Skipp and others, 1999). The mainly light to dark brown and gray, medium- to thick-bedded dolostones and dolomitic limestones of the Jefferson Formation (Dj) are prominently exposed in the center of the photograph. An increase in pale yellow and tan limestones and dolomitic siltstones near the top of the Jefferson is part of the Birdbear Member (Nisku; Dj_b). Some small folds and faults (orange dotted-dashed lines) are visible in the Jefferson. The entire Three Forks Formation (Dt) is ~50 m thick and exposed in the saddle just to the left (southwest) of the peak. The lower Logan Gulch member (Dt_l) conformably overlies the Jefferson Formation. The orange and tan, limonitic shales are topped by a ledge-forming limestone and dissolution breccia with variable thickness (collapse structures) and are visible near the saddle and peeking through the talus slope. The olive to green and gray mudstones of the Trident Member (Dt_t) are separated from the Sappington Formation (DMs) by another unconformity. Thin limestone beds of the Lodgepole Formation of the Mississippian Madison Group (Mm) are well exposed towards the top of the section. The black dotted rectangle is the location of fig. 18; the gray triangle refers to the same location in fig. 15B. Source of photo: author.

1.1 From Lithostratigraphy to Chronostratigraphy to Sequence Stratigraphy—The Kaskaskia Megasequence

The first geologic examination of Devonian strata in Montana was undertaken in the Three Forks area, northwest of Bozeman (fig. 2), in southwest Montana in 1860, by F.V. Hayden as a member of the United States Army mapping expedition let by William F. Raynolds. Subsequent visits to the Three Forks area by A.C. Peale and other members of the USGS in 1871, 1872, and 1884 provided the groundwork for the first geologic map of the area and the mention of Devonian strata in the State (fig. 4), namely the Upper Devonian Jefferson Formation and the Three Forks Shale (Peale, 1893). The occurrence of brachiopods and stromatoporoids in the Three Forks and Jefferson Formations was recognized early on and was critical for correlating Devonian strata into other parts of the State (e.g., Raymond, 1907, 1909; Haynes, 1916; Deiss, 1933, 1936, 1943; Berry, 1943; Sloss and Laird, 1946, 1947).

A general pinch-out of Devonian strata onto the Central Montana Uplift, an area of structural inversion during the Devonian (Woodward, 1996), limited the correlation of stratigraphic intervals across the State and resulted in the establishment of local stratigraphic names early on. Most infamous in this regard are the latest Famennian to earliest Tournaisian Bakken/Exshaw/Sappington Formation (figs. 3, 5). In the Milligan Canyon type section, west of Three Forks, Berry (1943) assigned the name Sappington Sandstone to ~18 m (60 ft) of yellow sandstone above the Three Forks Shale. Ten years later, Nordquist (1953) recognized a similar lithologic succession overlying the Three Forks Formation in the Williston Basin and named it the Bakken Formation. Another decade later, the Exshaw Shale, a black shale, sandstone, and carbonate unit in Canada (Warren, 1937), was formally described as extending from Canada into northwest Montana (Sandberg, 1966). This tripartite nomenclature for the latest Famennian and earliest Tournaisian strata is still used today, and all three names are recognized with formation status in Montana and used in the regions of their initial use.

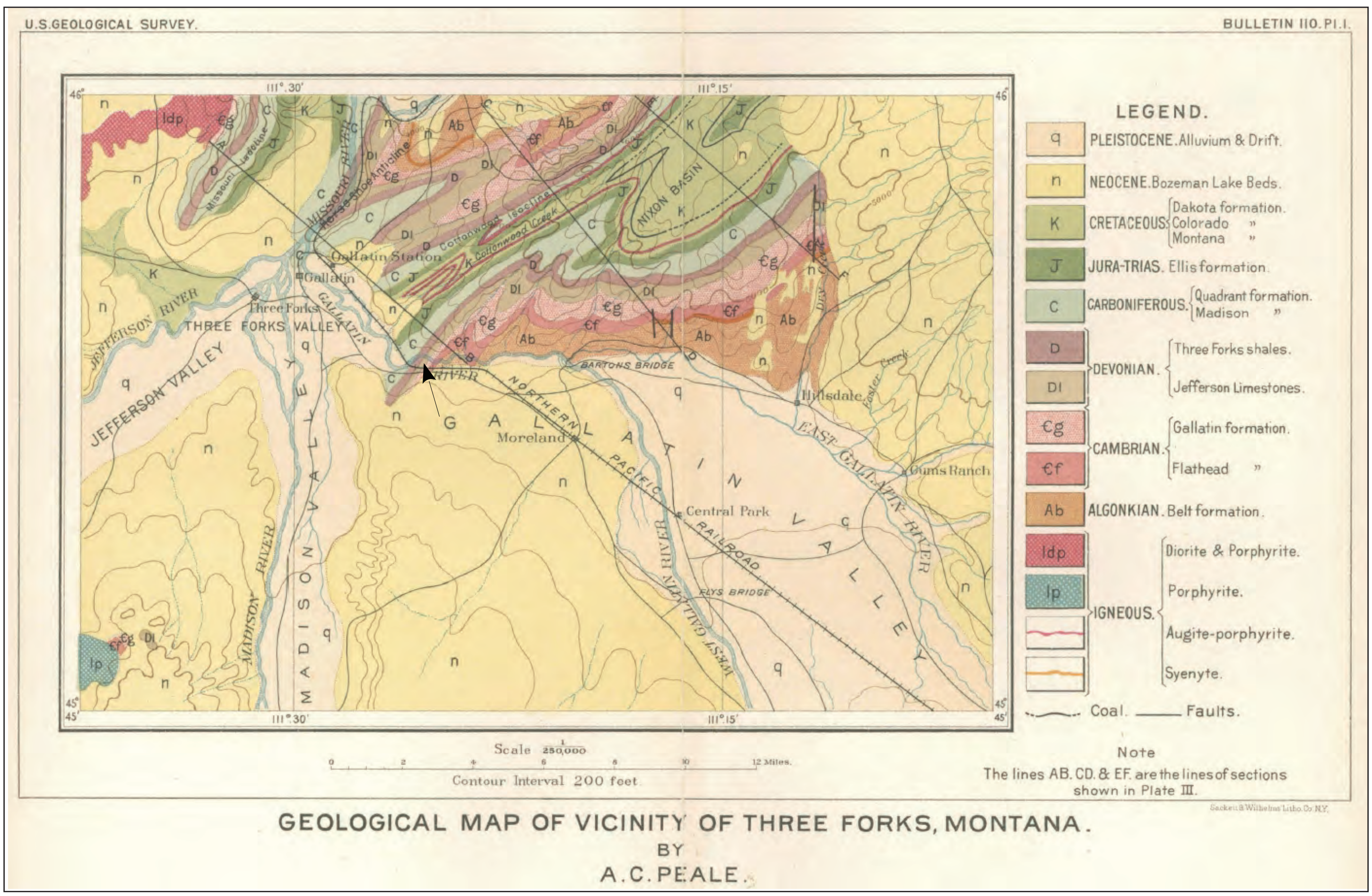


Figure 4. Copy of Peale's original 1893 geological map of the Three Forks area. Devonian stratigraphy was mapped for the first time in Montana. The sketch and photograph shown in figure 12 were taken at the present site of Logan, Montana (arrow). See also fig. 2 for location.

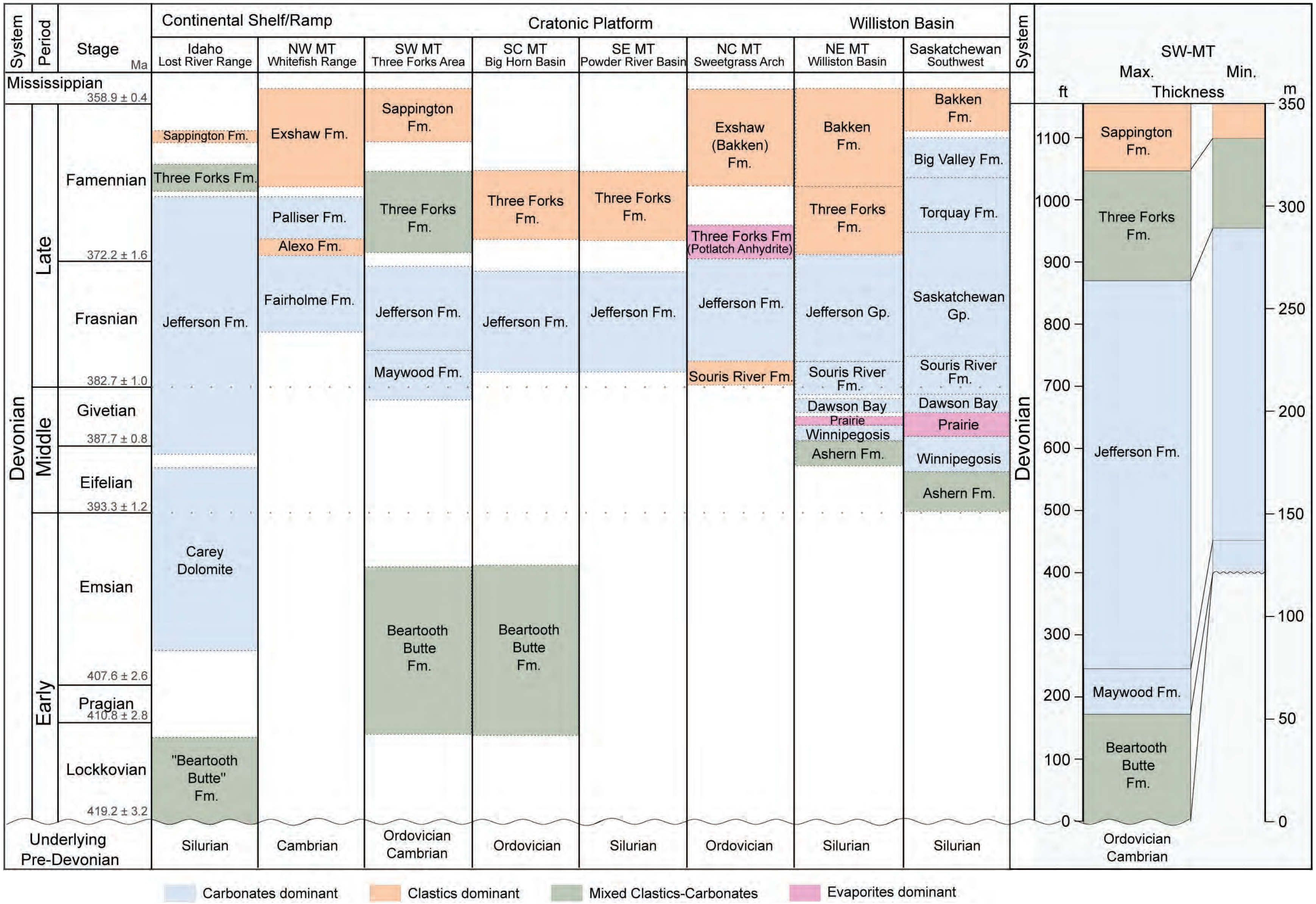


Figure 5. Chronostratigraphic correlation chart of Devonian strata (formation and group level) in Montana and adjacent areas. The most complete Devonian section in Montana is preserved in the Williston Basin (NE MT). Devonian strata generally thin towards central Montana onto the Devonian cratonic platform, then thicken to the west and into Idaho—the Devonian continental margin. The dominant facies throughout Montana are carbonates (blue), but clastics and mixed facies are present during the onset of Devonian deposition as well as during the waning stages. The carbonate facies dominance is best recognized when displaying the Devonian on a thickness scale (the two sections to the right with a light gray background), rather than on the time stratigraphic chart. Chart compiled with data from McMannis (1955), Sandberg (1961a, 1962a, 1965), Sandberg and Mapel (1967), Sandberg and McMannis (1964), Meyers (1971), Mallory and Hennerman (1972), Sandberg and Poole (1977), Balster (1980), Johnson and others (1985), Ehrets and Kissling (1987), Maughan (1989), Seward and Dyman (1990), Seward (1990), Grader and Dehler (1999), Stearns (2001), Schietinger (2013), Saskatchewan Ministry of the Economy (2014), Grader and others (2014, 2016), Rodriguez and others (2016), diPasco and others (2017), Hogancamp and Pocknall (2018).

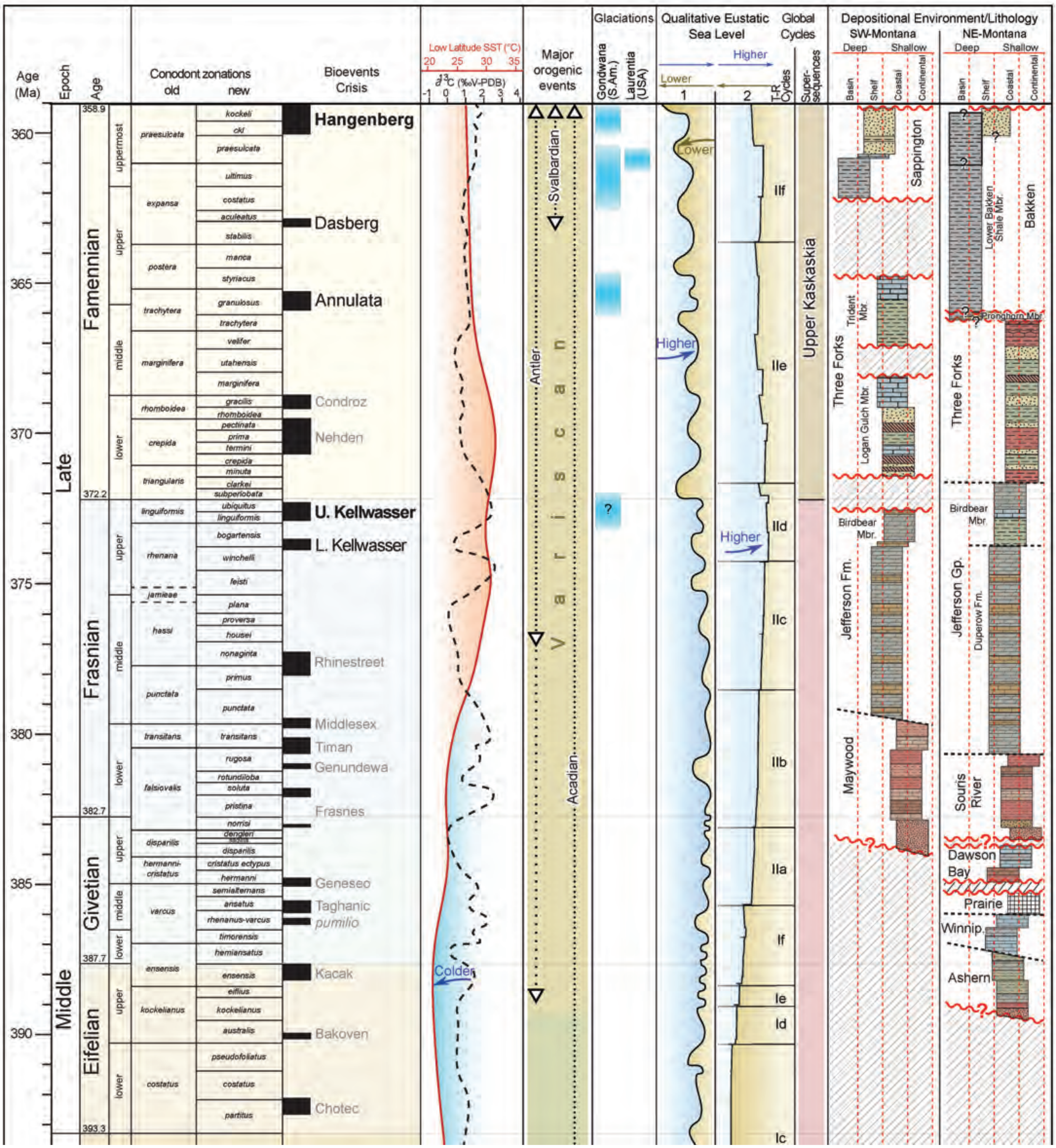
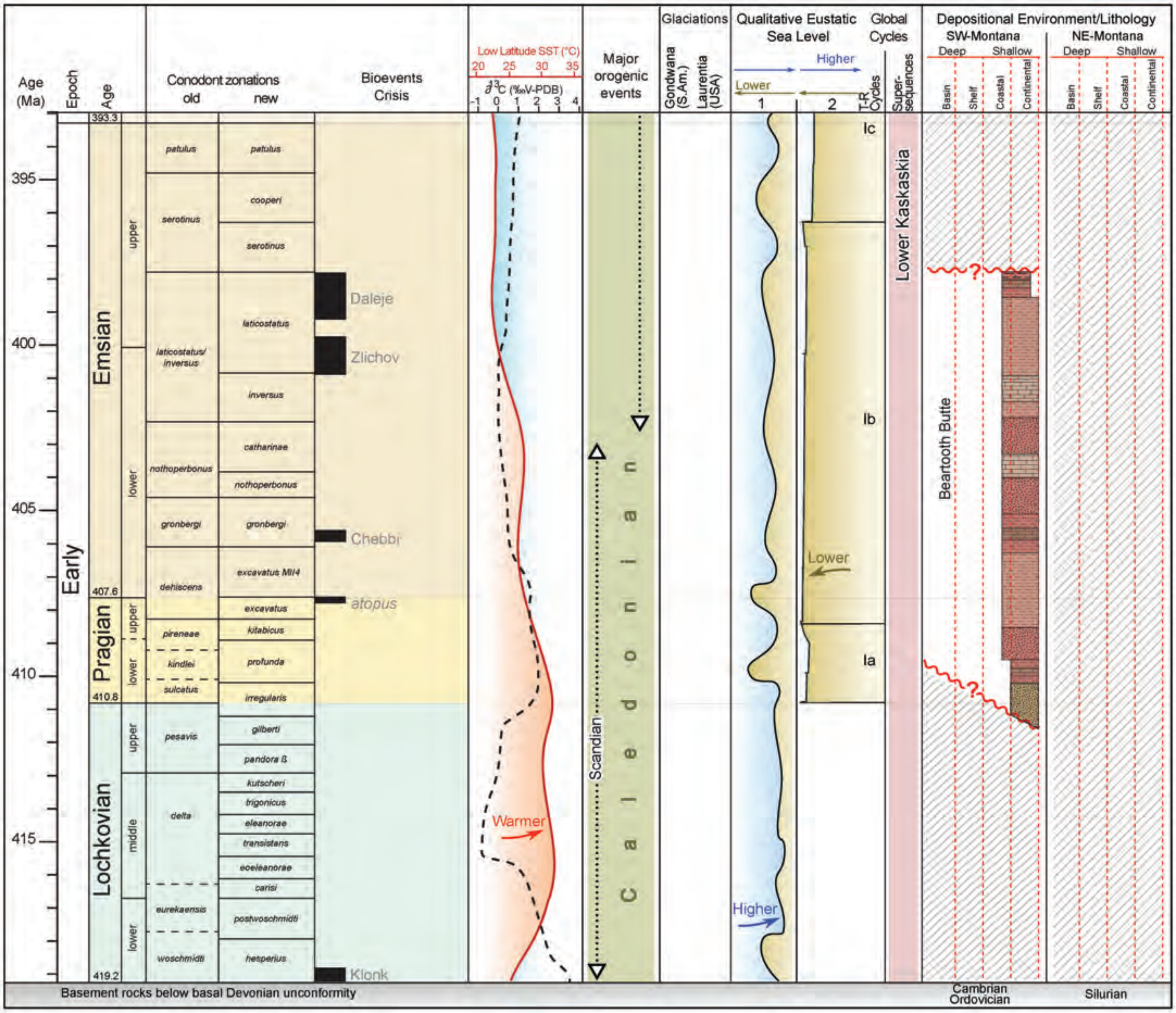


Figure 6. Compilation chart of global and local Devonian events and stratigraphy. From left to right, the tracks are chronostratigraphic scale (age) in million years (Ma), from Becker and others (2016) and Brett and others (2020); epoch and age boundaries scaled to chronostratigraphic scale; conodont zones (old and new) scaled to chronostratigraphic scale, from Becker and others (2016); bioevents (1st order bioevents in bold text, 2nd order in regular text, 3rd and higher order in gray text), modified from Becker and others (2016), and Brett and others (2020); low-latitude sea surface temperature in °C (SST; red solid line) and carbon isotope data in permill ($\delta^{13}C$; black dashed line), modified from Buggisch and Joachimski (2006) and Joachimski and others (2009); major tectonic events (global and relevant to Montana, double arrows and color bars) based on data from Dorobek and others (1991), and McKerrow and others (2000); Devonian glaciation events in Gondwana and Laurentia (blue boxes) reported in Streel and others (2000), Caputo and others (2008),

(Caption continued on next page)



Isaacson and others (2008); qualitative eustatic sea level, 1 from Haq and Schutter (2008), and 2 from Johnson and others (1985); T-R cycles (Johnson and others, 1985) and Supersequences (Haq and Schutter, 2008) scaled to the chronostratigraphic scale; stratigraphic column, including dominant lithology and major unconformities, for southwest Montana (fig. 2 for location) and northeast Montana, scaled to the chronostratigraphic scale (vertical scale), and depositional environment (horizontal scale). Montana stratigraphic columns compiled with data from McMannis (1955), Sandberg (1965), Sandberg and McMannis (1964), Sandberg and Mapel (1967), Meyers (1971), Mallory and Hennerman (1972), Sandberg and Poole (1977), Ehrets and Kissling (1987), Stearns (2001), Grader and others (2014, 2016), Rodriguez and others (2016), di Pasco and others (2017), Hogancamp and Pocknall (2018).

The recognition of cratonal correlative megasequences (Sloss, 1963) followed by the development of seismic and sequence stratigraphic principles (Payton, 1977; Vail and others, 1977; Vail, 1987) revolutionized how the stratigraphic record is analyzed. The subdivision of lithologic successions into related depositional sequences, bound by unconformities and other significant surfaces, provides a useful tool for analyzing genetically and chronologically related stratigraphy and facies distributions in an area. The Devonian in Montana is no exception, because the base of the Devonian is marked by a regional unconformity, the basal Devonian unconformity, that is onlapped by the time-transgressive strata of the Kaskaskia Megasequence, the third oldest of the six cratonic megasequences recognized in North America (fig. 6, previous pages; Sloss and Laird, 1947; Sloss, 1950, 1963; Sandberg and others, 1988). The Devonian rocks in western Montana overlie Cambrian strata, in central Montana the Devonian strata is in contact with Ordovician rocks below the unconformity, and in the northeastern part of the State, Silurian strata is truncated by the basal Devonian unconformity (fig. 5).

In general, the Kaskaskia Megasequence in Montana can be separated into two distinct higher order supersequences. The lower Kaskaskia (Kaskaskia I) Supersequence (fig. 6) is of mid-Early Devonian to latest Devonian (Pragian–Famennian) age, and the upper Kaskaskia (Kasakskia II) Supersequence is latest Devonian (Famennian) to late Mississippian (Visean/Serpukhovian; Haq and Schutter, 2008). Carbonate rocks are the dominant lithology (by thickness) in the Kaskaskia Megasequence rock record in Montana, but evaporites and siliciclastic deposits also occur to a lesser degree, the latter in particular during the onset and waning stages of deposition of Devonian strata (figs. 5, 6). This Devonian review largely focuses on the deposits of the Kaskaskia I Supersequence and the early stages of the Kaskaskia II Supersequence (Famennian) that are described in ascending temporal order, starting from the oldest Devonian rocks that are recognized only locally in western Montana, to the youngest formations that bridge the Devonian to Mississippian boundary throughout Montana.

STRATIGRAPHY, FACIES, AND FACIES DISTRIBUTION

Early and Middle Devonian

The oldest Devonian deposits preserved in Montana belong to the Beartooth Butte Formation (figs. 5–7), a heterolithic accumulation of thin-bedded red and buff mudstone, gray to yellowish carbonate mudstone, sandstone, sandy and silty dolomite, limestone conglomerate and breccia, and light gray to grayish red dolomite containing fish and plant fossil remains (Dorf, 1934; Sandberg, 1961a; Sandberg and Mapel, 1967; Meyers, 1971; Fiorillo, 2000). In outcrop the formation is only found locally in southwestern, south-central, and central Montana, as far north as the Big Snowy Mountains (fig. 2). Where it occurs, its thickness varies over short distances but can be as much as 52 m (170 ft; fig. 7; Sandberg, 1961a; Sandberg and Mapel, 1967; Meyers, 1971).

The North American continent remained largely exposed during the Middle Devonian, and deposition was limited to isolated basins (fig. 8A). The same is true in Montana, where the onset of more widespread deposition started in the Middle Devonian, although first only in eastern Montana (fig. 8B). There, the onset of Kaskaskia transgression is marked by the siliciclastic and mixed carbonate siliciclastic units of the Ashern Formation (figs. 5, 6; Baillie, 1951). The thin Ashern Formation was included as part of the Winnipegosis Formation (Sandberg and Hammond, 1958), but others recognized it as the basal formation of the Elk Point Group (McGehee, 1949; Belyea, 1952; Baillie, 1953, 1955; Sandberg, 1961b; Lobdell, 1984). Breccias in the lower parts (lower member after Lobdell, 1984) of the Ashern Formation are interpreted as transgressive lags and contain abundant reworked Silurian subcrop strata. Claystone, silty and argillaceous dolomite, dolomitic shale, and dolomitic limestone are common facies; evaporites occur locally in the lower Ashern Formation, whereas cephalopods, gastropods, and brachiopods are more common near the top of the formation. In general, lithologies in the lower Ashern Formation have a more intense red color and change to a greener hue higher up in the succession (Baillie, 1953; Lewis, 1958; Lobdell, 1984; Rosenthal, 1987; Megathan, 1987).

Overlying the mixed clastic-carbonate facies of the Ashern Formation are mainly limestone beds of the Winnipegosis Formation (figs. 5, 6; Tyrrell, 1892).

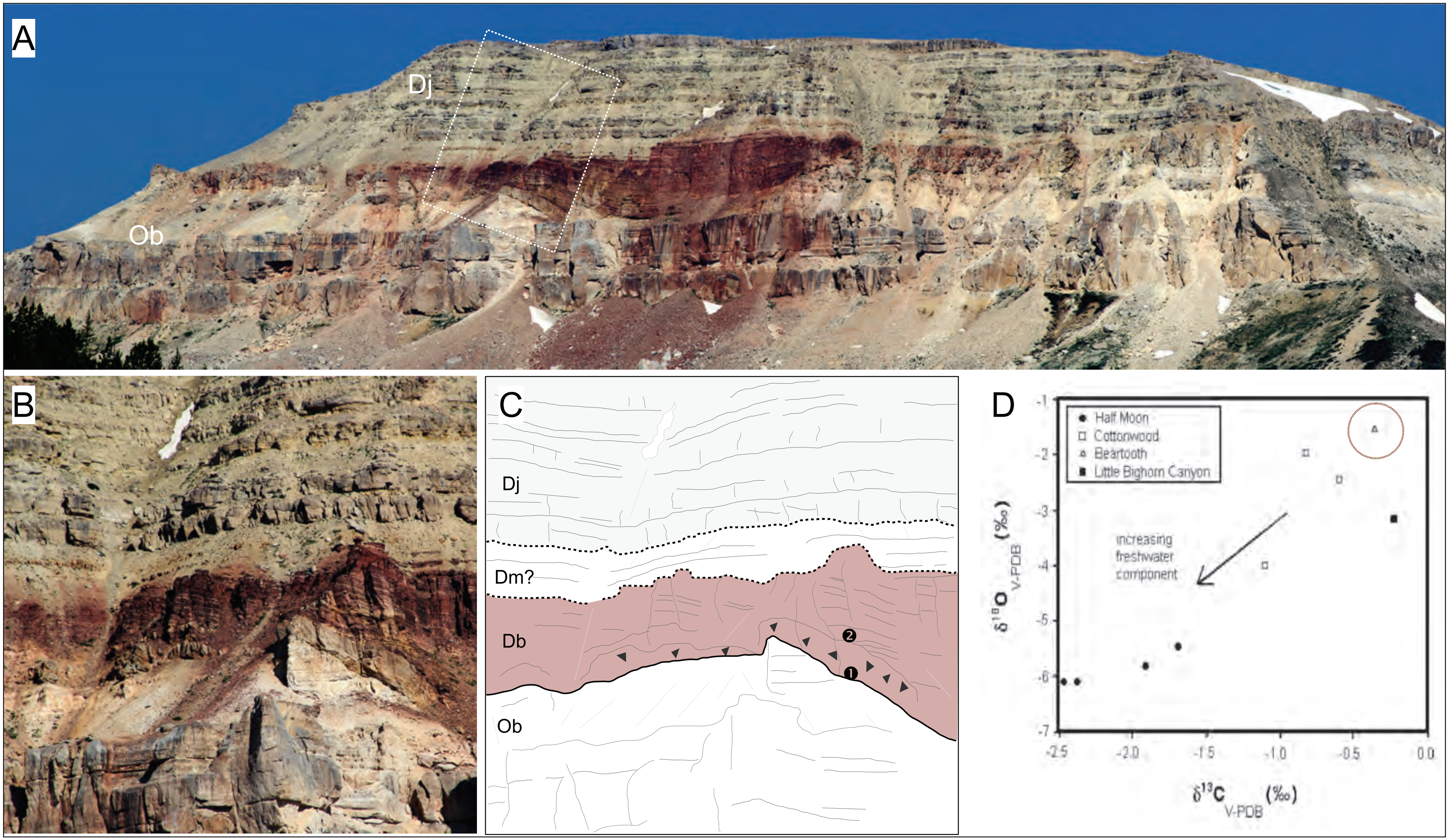


Figure 7. The Beartooth Butte Formation at its type locality at Beartooth Butte, WY, in the Beartooth Mountains, just ~3.5 mi south of the Montana border (see fig. 2 for location). (A) Beartooth Butte viewed from “Top of the World” on Beartooth Highway (Hwy 212). The intensely red Beartooth Butte Formation is infilling channel-like incisions carved into the underlying, largely gray Ordovician Big Horn Dolomite (Ob; Sandberg, 1961a). Subsequently, the Beartooth Butte Formation was overlain by the nearly horizontal beds of the younger Devonian strata. Most recognizable are the beds of the Frasnian Jefferson Formation (Dj). View is due west; height of outcrop is ~800 ft. Thickness of Beartooth Butte Formation is approximately 150 ft (50 m; Fiorillo, 2000). Close-up photograph (B) and line drawing (C) of the Beartooth Butte Formation from Beartooth Lake (view is due NW; see white dashed outline in A for approximate photo area). Conglomerates (triangles in C) form the basal facies of the Beartooth Butte Formation. The thin-bedded red mudstones onlap the basal conglomerates. The inclined beds gradually flatten upward. Horizontal, tan to yellowish, very thin beds can be seen atop the red Beartooth Butte thin beds and might be deposits of the Maywood Formation (Dm). The medium-bedded, medium gray beds above are the dolomite facies of the Frasnian Jefferson Formation (Dj). (D) Plot of stable oxygen and carbon isotope data from Beartooth Butte Formation localities in Montana and Wyoming, showing the increase in freshwater influence from the Beartooth Butte type locality towards the north into central Montana (Half Moon samples; figure from Fiorillo, 2000). Source of photos: author.

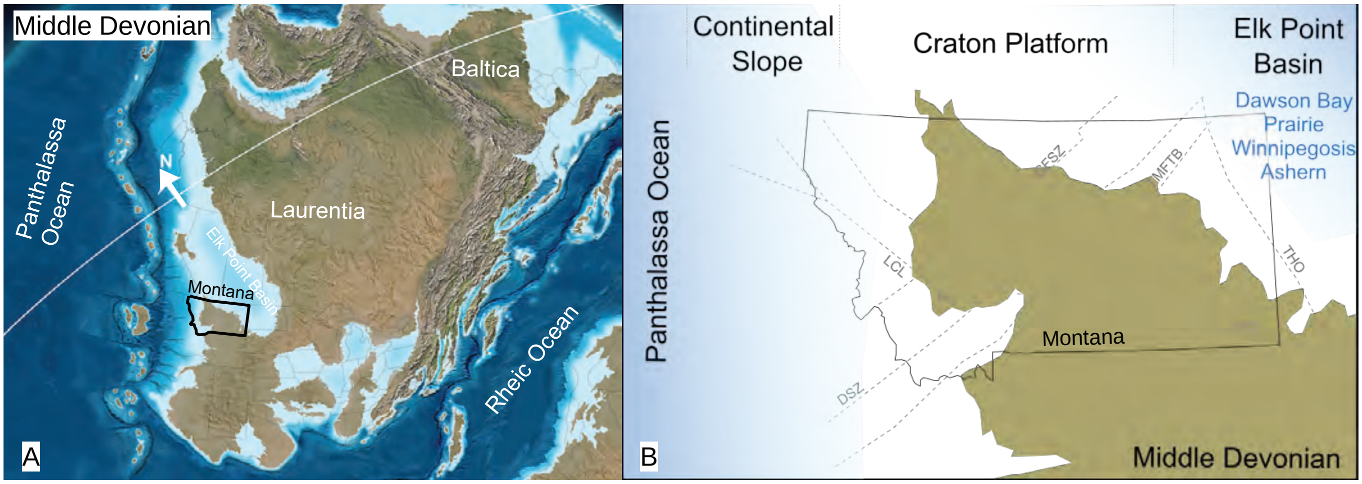


Figure 8. Middle Devonian paleogeography. Baltica and Laurentia were almost fully exposed during this time and the Kaskaskia transgression only inundated isolated areas, including the Elk Point Basin in northeast Montana (A). Early Middle Devonian rocks (Eifelian) are only found in the subsurface in northeastern Montana, including the Ashern, Winnipegosis, Prairie, and Dawson Bay Formations (B). During the late Middle Devonian (Givetian), continued flooding of the craton from the west and the east resulted in more widespread deposition of the Maywood and Souris River Formations (see also fig. 10). Basement lineaments (gray dashed lines) might have had some control on the paleogeography, but a clear trend is not obvious during this time in part because of the limited exposures of these deposits. DSZ, Dillon Shear Zone; GFSZ, Great Falls Shear Zone; LCL, Lewis and Clark line; MTL, Mesoproterozoic Montana–Tennessee line; PL, Perry line; THO, Trans-Hudson orogen; TMFTB, Trans-Montana fold-and-thrust belt. Maps compiled and modified from Blakey (2016), and Sims and others (2004).

The thickness of the Winnipegosis Formation in the Williston Basin ranges from 0 to 122 m (0 to 400 ft; Baillie, 1953, 1955; Sandberg, 1961b), and the unit is mainly composed of dolomitic shale, dark gray calcareous shale and siltstone, argillaceous dolomite, light gray to brownish gray dolomite, and fossiliferous limestone (fig. 9; Sandberg and Hammond, 1958; Sandberg, 1961b; Perrin, 1982, 1987; Ehrets and Kissling, 1987).

The youngest formation of the Elk Point Group is the Prairie Formation (figs. 5, 6). In Montana, the Prairie Formation only occurs in the subsurface in the far northeastern corner of the State, mainly in Sheridan County, where it thickens to over 61 m (200 ft; Nicolas, 2015), consisting of mainly halite (fig. 9), with lesser sylvite, potash, anhydrite, dolomite, clay, and quartz (Berg, 2010). The facies changes laterally, including interbedded anhydritic dolostones, dolomitic mudstone, and siltstone with halite inclusions, and can reach a thickness of 183 m (600 ft) in the central parts of the Williston Basin in North Dakota and Saskatchewan (Sandberg and Hammond, 1958; Bannatyne, 1983; LeFever and Le Fever, 2005; Nicolas, 2015).

Overlying the Elk Point Group is the Dawson Bay Formation (figs. 5, 6). Like the Elk Point Group, the Dawson Bay Formation is only described in the subsurface in northeastern Montana. The Dawson Bay Formation overlies the Prairie Formation—the basal contact of the Dawson Bay Formation is described as conformable by some, but unconformable by others (e.g., Sandberg and Mapel, 1967; Ehrets and Kissling, 1987)—and is composed of fine-grained siliciclastic rocks, argillaceous limestone and dolomite, and gray red silty argillaceous dolomite (Baillie, 1953; Sandberg and Hammond, 1958). These mostly siliciclastic basal deposits are succeeded by massive dolomitic limestone and limestone with corals and stromatoporoids, and local anhydrite and anhydritic dolostone in the upper parts (fig. 9; Baillie, 1953; Sandberg and Hammond, 1958). Anhydrite can fill small vugs in brecciated limestones and dolostones, together with calcite (fig. 9).

In contrast to the majority of the Middle Devonian strata that are largely only present in the subsurface in northeastern Montana, Late Devonian strata are widespread across the State (fig. 5). The first deposits of very late Middle Devonian (late Givetian) to early Late Devonian (early Frasnian) age are the Souris River Formation (figs. 5, 6)—present in the subsurface in eastern Montana (including the Williston Basin), in the central Montana uplift, the Sweetgrass Arch, and in the Big Horn Basin of south-central Montana—and the time-equivalent Maywood Formation in western and southwestern Montana (late Givetian to early Frasnian; Sandberg, 1962a, 1967; Kauffman and Earll, 1963; Mayers, 1971).

The thickness of the Maywood and Souris Formations varies greatly (McMannis, 1962; Sandberg and

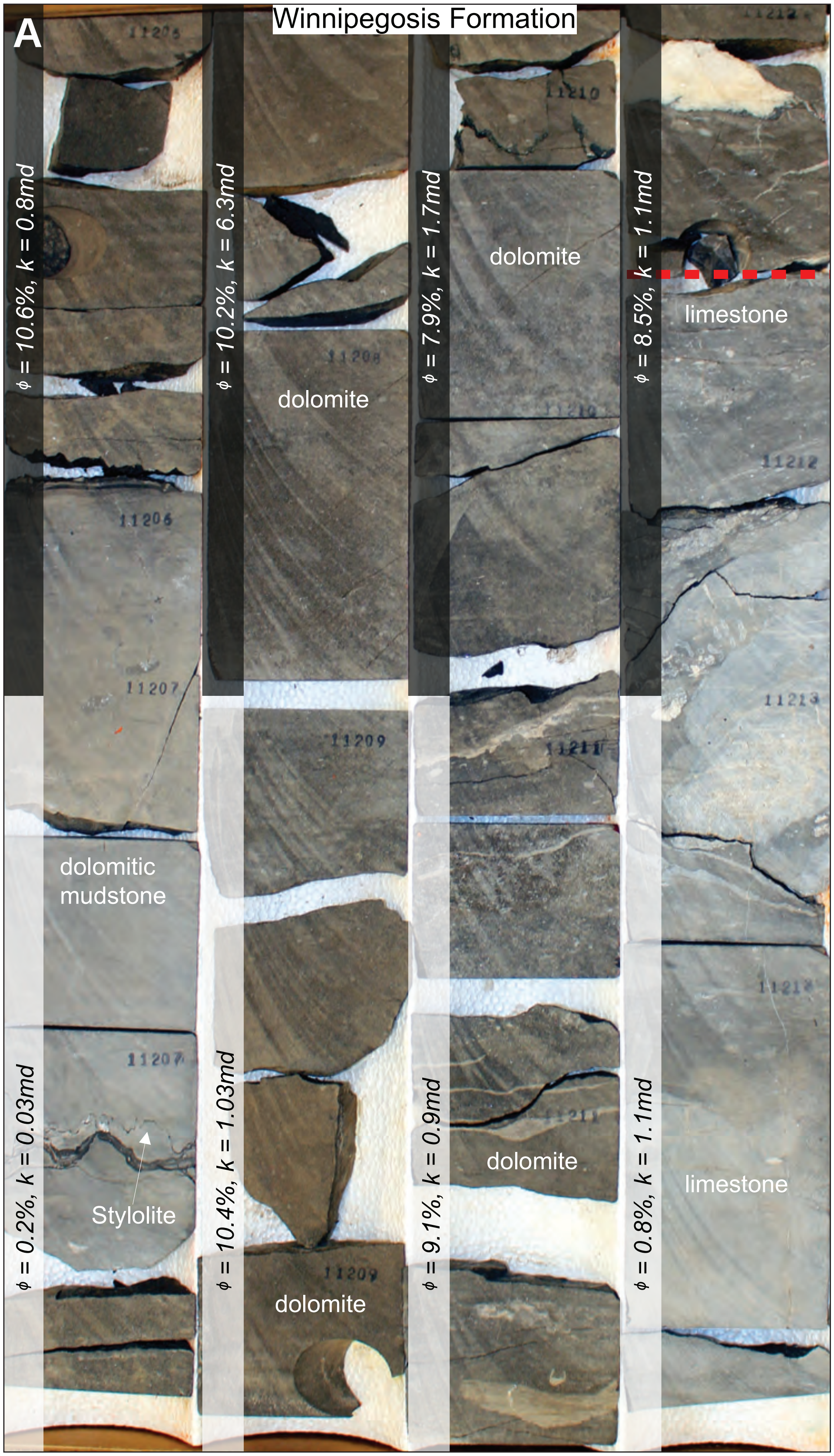


Figure 9A. Middle and early Late Devonian facies in core. This example from the western Williston Basin (Roosevelt County) shows the heterolithic Winnipegosis facies compressed in one core box with Helium porosity and maximum air permeability listed alongside. Mainly dark gray and brownish gray sucrosic dolomite (above red line) and thin interbeds of dolomitic mudstone (at 11,207 ft marker). Apart from the mudstone interval, the porosity of the dolomites is consistently at approximately 10%, although permeability varies from <1 md to >6 md. A centimeter-scale amplitude stylolite at the base of the mudstone interval provides evidence for abundant pressure solution. Below the dashed red line, well-cemented (low porosity and permeability), brecciated, and fossiliferous limestones are the dominant facies. Photo from core Granley State 4-15X (API#: 2508521604), Roosevelt County, MT. Core box height is 2 ft. Photo source: author.

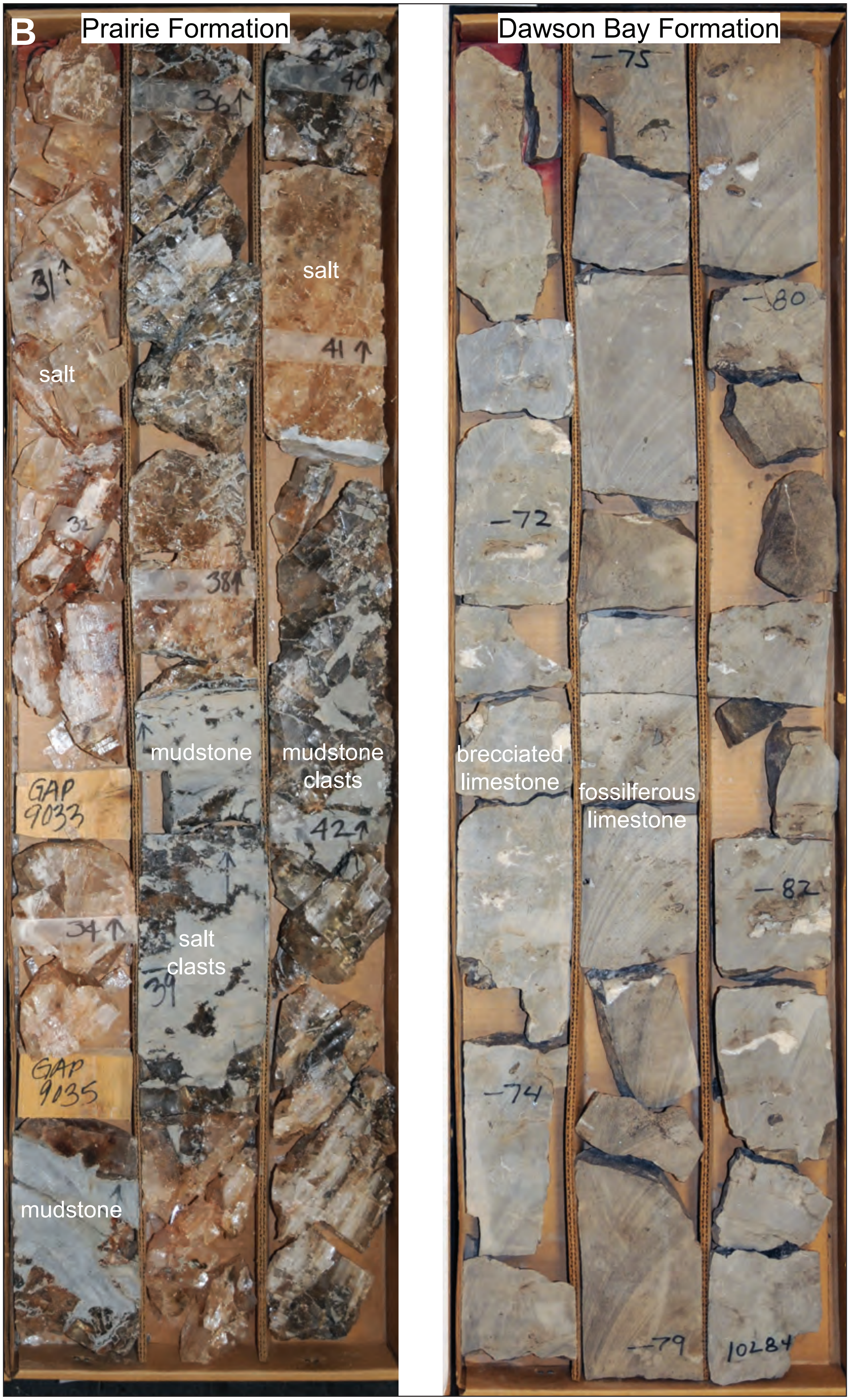


Figure 9B. Middle and early Late Devonian facies in core. The Prairie Formation (left) contains abundant salt deposits interbedded with intervals of mudstones. Photo from core W.H. Hass #1 (API#: 2509121328), Sheridan County, Montana. Core box height is 3 ft. Photo source: USGS, https://my.usgs.gov/crcwc/. This example of mainly calcareous and dolomitic facies of the Dawson Bay Formation (right) is from Dawson County, Montana. Photo from core Burlington Northern 1 (API#: 2502121041), Dawson County, Montana. Core box height is 3 ft. Photo source: USGS, https://my.usgs.gov/crcwc/.

McMannis, 1964; Grant, 1965; Meyers, 1971, 1980; Thomas, 2011), but generally increases from central Montana to the west towards the Devonian shelf margin and northeast into the Williston Basin (figs. 5, 10; e.g., Emmons and Calkins, 1913; Sandberg, 1962a; Kauffman and Earll, 1963; Carlson and Anderson, 1965; Witkind, 1971; Mudge, 1972; Mudge and others, 1982; Porter and Wilde, 2001), with the greatest thickness of about 120 m (394 ft) reported from northwest Montana (Vuke and others, 2007). In central and southern Montana, the Maywood and Souris River Formations are absent; however, the position of this pinch-out is not very well constrained. Figure 10 highlights the area of uncertainty for the zero edge of the Maywood/Souris River Formations and also illustrates the ongoing transgression of the Kaskaskia seas onto the craton (compare to paleogeography in fig. 8).

The facies of the early Middle Devonian strata are heterolithic. In some areas in western and eastern central Montana, the basal facies are thin conglomerates (Sandberg and Hammond, 1958; Sandberg, 1961b; Thomas, 2011). More common are tan, brown, yellow, red, greenish gray, and green, thin-bedded to laminated, finely crystalline limestone, dolomitic limestone, microcrystalline and finely crystalline dolostone, silty dolostone, argillaceous dolostone, dolomitic mudstone, and thin beds or laminae of siltstone, siliciclastic mudstone, and calcareous shale (Emmons and Calkins 1913; Meyers, 1971; Witkind, 1971; Mudge, 1972; Porter and Wilde, 2001; Vuke and others, 2014; Thomas, 2011). Siliciclastic interbeds (figs. 11A–11C) are most common in central Montana (Witkind, 1971), and also generally increase towards the north (fig. 10) and northeast (Sandberg and Hammond, 1958; Sandberg, 1961b; Carlson and Anderson, 1965; Porter and Wilde, 2001). Anhydrite beds are only present in the Souris River Formation in eastern Montana (fig. 6; Sandberg and Hammond, 1958; Sandberg, 1961b).

Late Devonian

Where present, the Maywood and Souris River Formations are conformably overlain by the Jefferson

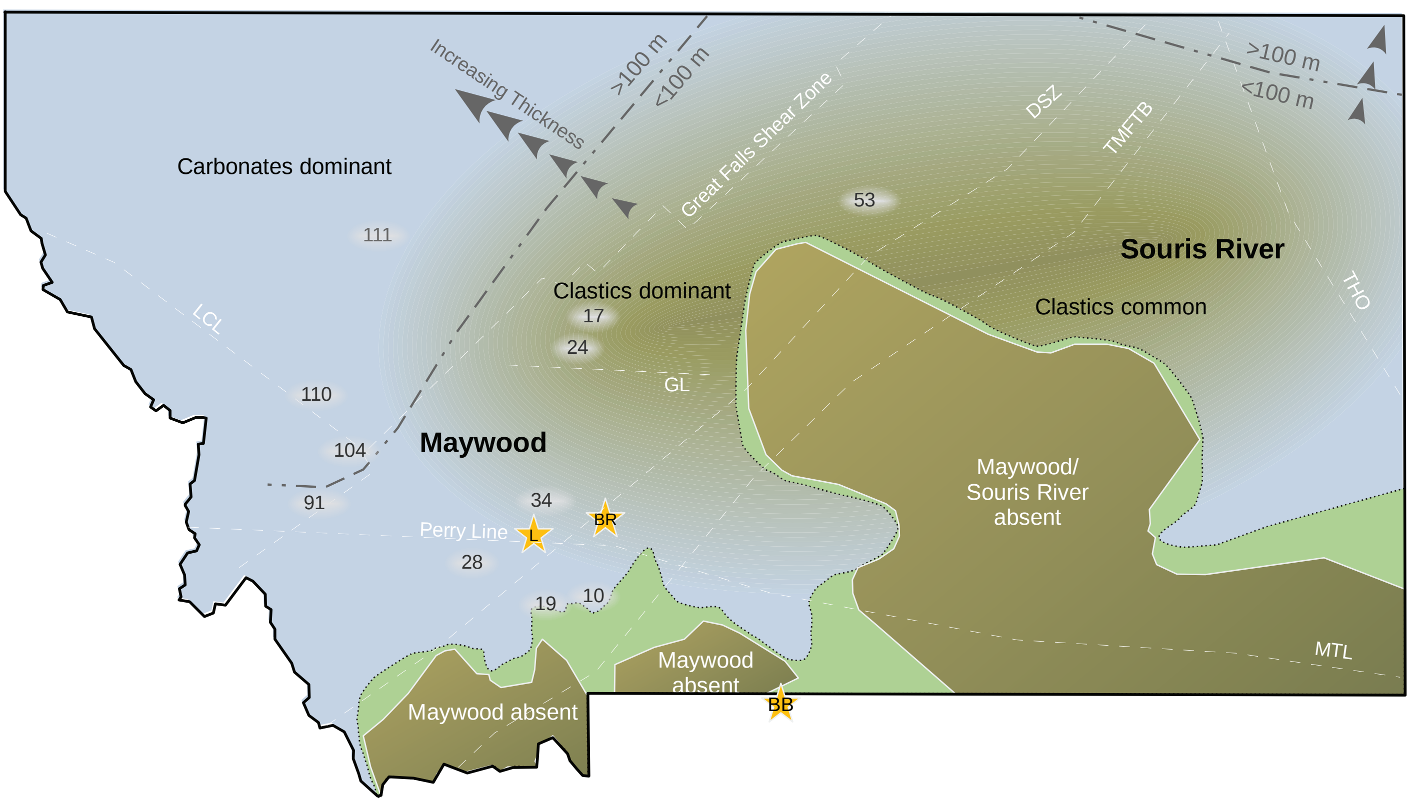


Figure 10. Map of the Maywood and Souris River Formations in Montana. The position of the depositional limit and/or erosional edge of the Maywood and Souris River Formations has been mapped at different locations throughout the years. The pinch-out envelope (gray) marks the zone of uncertainty gathered from these reports. Data to constrain the pinch-out envelope from Sandberg and Hammond (1958); Sandberg and McMannis (1964); Carlson and Anderson (1965); Grant (1965); Benson (1966); Meyers (1971); Poole and Sandberg (1977); and Thomas and others (1996). The 100 m contour line and arrows show this general thickness trend; the small black numbers are thickness in meters. The facies of the Maywood and the Souris River Formations is variable; carbonates are most common (blue colors); clastic deposits are dominant locally (dark yellow color). The stars mark locations of outcrop photographs shown in this report. L, Logan; BR, Bridger Range; BB, Beartooth Butte. White dashed lines are basement lineaments (from Sims and others, 2004).

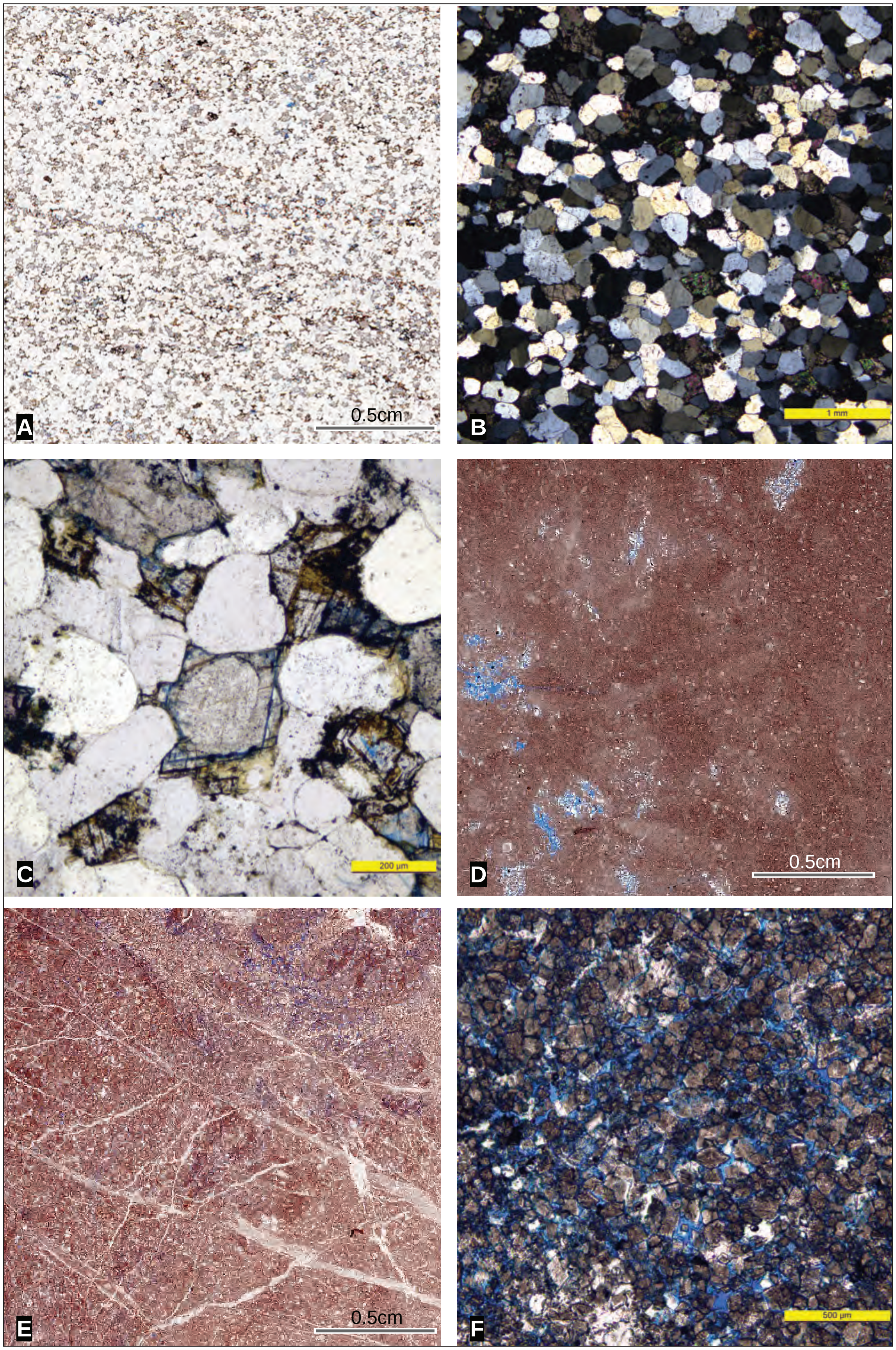


Figure 11. Facies examples of the Maywood Formation (A–C) and the Jefferson Formation (D–L) from Mulkey Gulch in west-central Montana and of the Jefferson Formation from the southern Pioneer Mountains (G–H) and from Clark Canyon Reservoir (I–L) in southwest Montana (see fig. 2 for location). Thin section scan (A) and cross-polarized light photomicrograph (B) of fine- to medium-grained quartzarenite in the Maywood Formation. The well-rounded grains have widespread quartz overgrowth (B, C), and euhedral dolomites fill some intergranular pores and replace grains locally (C). The Jefferson Formation contains mottled, finely crystalline dolomite, with

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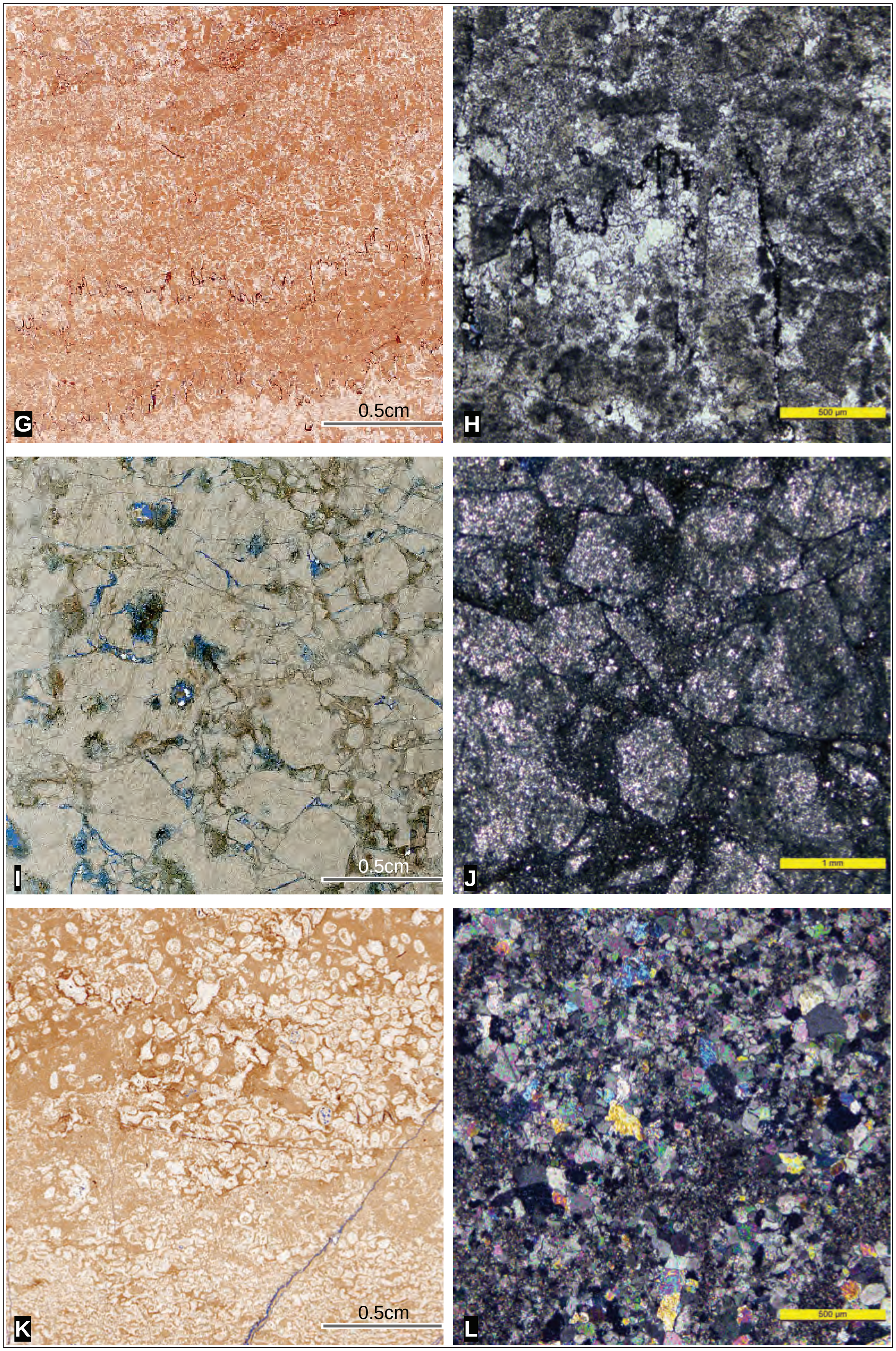

large vugs (D). Some vugs are calcite cemented (white); others remain open (blue). Some of the dolomites are heavily fractured (E). Fractures are commonly calcite cemented (white). Calcite is also present as patchy cement in intercrystalline pores; however, porosity is well preserved between the finely crystalline dolomite crystals (F). Stylolites are evidence for pressure solution after dolomitization (G, H). Breccias (I, J) and patchy dolomitization of calcareous, fossiliferous limestones (K, L) are common in the Jefferson. Photo source: author.

Group [including the Duperow and Birdbear (Nisku) Formations] in eastern Montana, the Jefferson Formation [including the Lower (Duperow) Member and Birdbear Member] in western Montana, and the Fairholme Group, Alexo Formation, and lower Palliser Formation in northwestern Montana (figs. 5, 6). The name Jefferson Limestone was originally assigned to fossiliferous, brown and black crystalline limestones near the town of Three Forks, but an official type section was not established (Peale, 1893). Subsequently, the Jefferson type section (fig. 12) was established north of the Gallatin River, across the river from Logan, Montana (Sloss and Laird, 1946, 1947; Sandberg, 1962a), where the Jefferson Formation has a total thickness of ~165 m (540 ft; Sandberg, 1965). The thickness of these time-equivalent stratigraphic units varies significantly locally and across the State, with the thinnest strata [<30 m (100 ft)] on the Central Montana Uplift and the Beartooth Wyoming Shelf (McMannis, 1962; Lindsey, 1980; Peterson, 1984; Dorobek, 1991; Porter and others, 1996) and a general thickening to the west (fig. 13); the Jefferson Formation is >518 m (1,700 ft) thick in the Garnet Range in western Montana (Kauffman and Earll, 1963) and continues to thicken into Idaho (fig. 5), and the Jefferson Group is as much as 221 m (725 ft) thick in the Williston Basin in northeastern Montana (Sandberg and Hammond, 1958).

The Jefferson Formation/Group and time-equivalent strata (fig. 5) contain a richly diverse fauna (Cooper and others, 1942; Berry, 1943; Sloss and Laird,

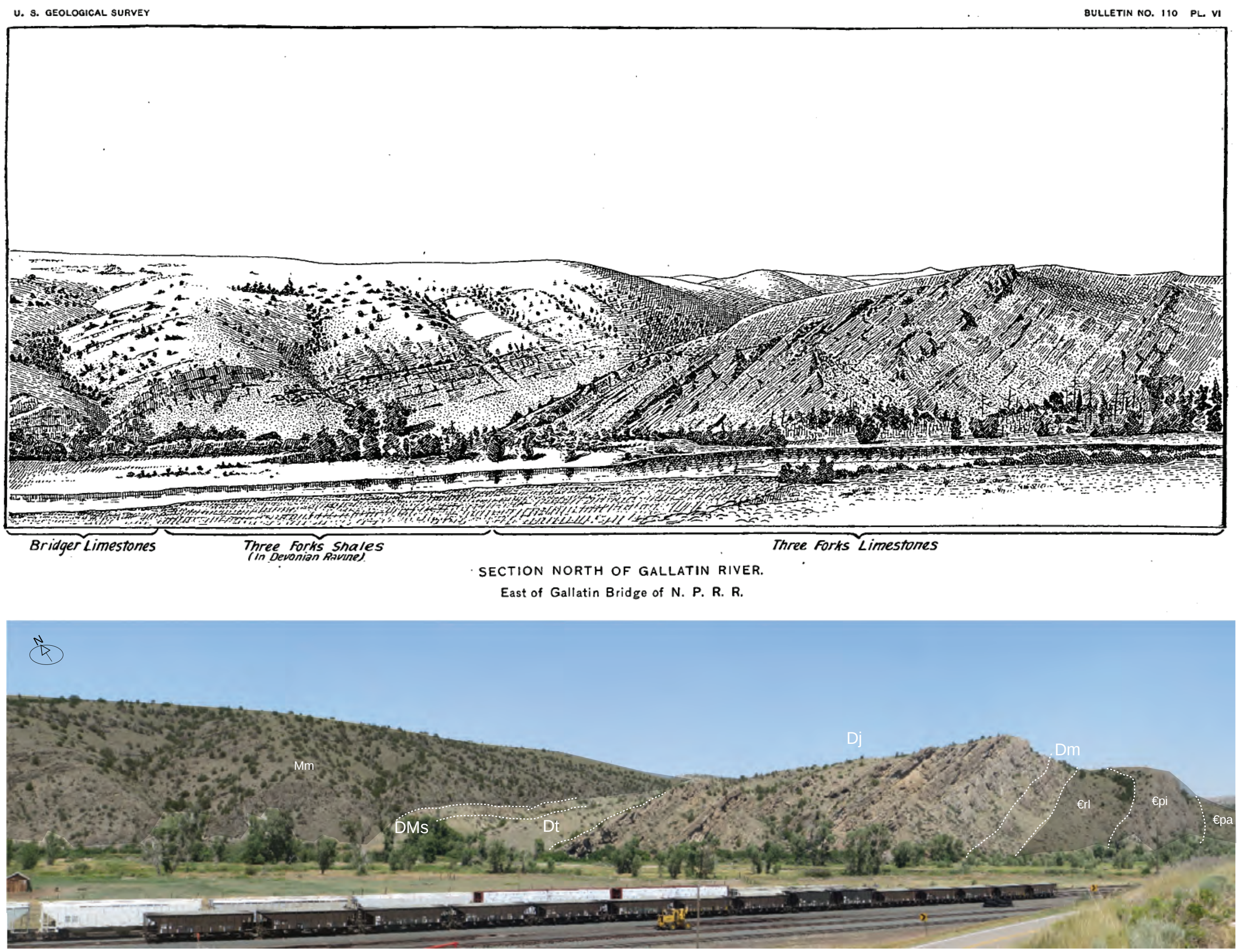

Figure 12. The Devonian type section near Logan, Montana. Peale's original sketch (A) of the Devonian outcrops north of the Gallatin River and a photograph (B) of the same outcrops north of Logan, MT. The Three Forks Limestones from Peale's 1893 paper include Cambrian Park Shale (€pa), Pilgrim Limestone (€pi), Red Lion Sandstone (€rl), Devonian [Maywood (Dm)], and Jefferson (Dj) strata. The Three Forks Shale (in "Devonian Ravine") of Peale are equivalent to the Three Forks Formation (Dt) and the Sappington Formation (Ds), respectively. The Bridger Limestones are equivalent to the limestones of the Mississippian Madison Group (Mm). For location see figs. 2, 4. Photo B source: author.

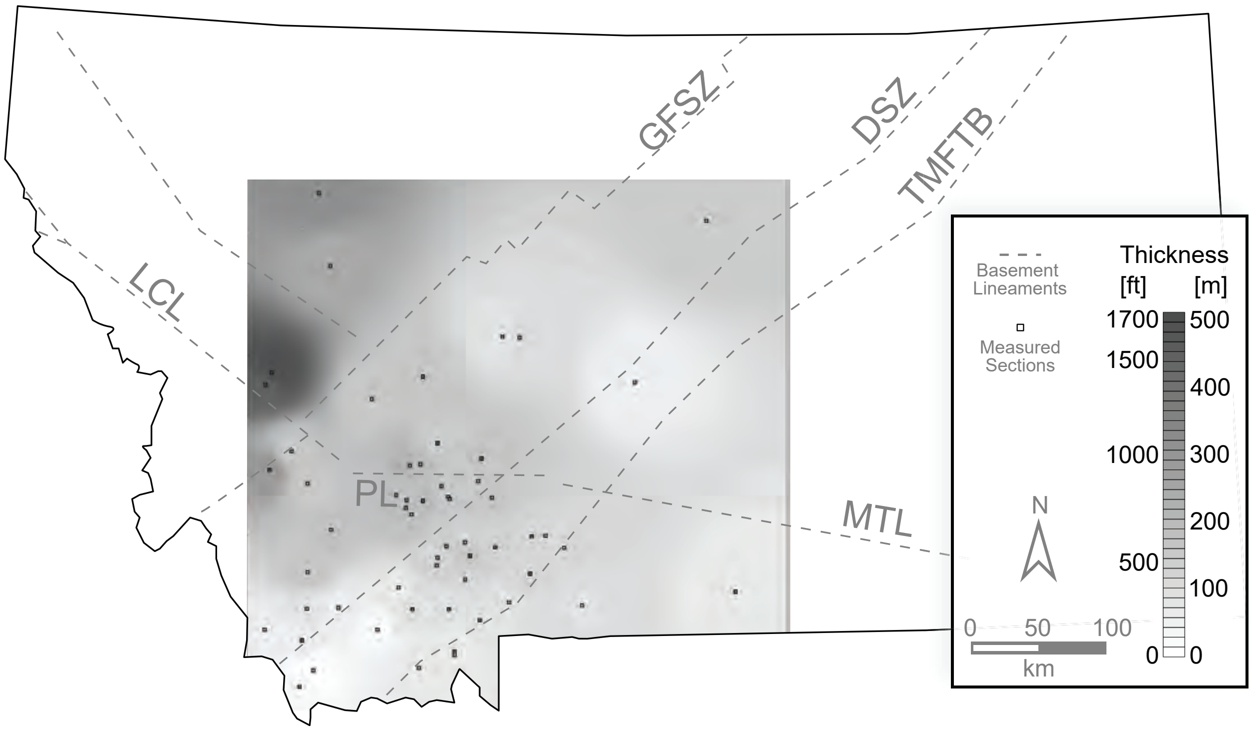

Figure 13. Map of Jefferson thickness (ft) from outcrop measurements in southwest and central Montana. The Jefferson Formation generally thickens to the west (dark gray areas) and is thinnest to the southwest and towards central Montana (light gray areas). The thinnest Jefferson thickness, located in SW Montana, is bound by Paleoproterozoic Sutures (gray dashed lines), namely the Trans-Montana fold and thrust belt (TMFTB) to the southeast and the Dillon Shear Zone (DSZ) to the northwest. Other notable thickness changes occur just northeast of the Lewis and Clark Line (LCL) and northwest of the Great Falls Shear Zone (GFSZ). Jefferson thickness data (squares) from Deiss (1933), Sloss and Moritz (1951), McMannis (1955), Richards (1955), Klepper and others (1957), Sandberg (1961b), Kauffman and Earl (1963), Knopf (1963), Robinson and Barnett (1963), McMannis and Chadwick (1964), Sandberg (1965), Benson (1966), Witkind (1971), Mudge (1972), Lindsey (1980), Kellogg (1992), Porter and others (1996), Skipp and others (1999), Lopez (2000), Porter and Wilde (2001), Vuke and others (2002), Thomas (2011), Stickney and Vuke (2017), and Lund (2018). Basement structural lineaments from Sims and others (2004).

1947; Nave, 1952; Berg, 1959; McMannis, 1962; Robinson and Barnett, 1963; McMannis and Chadwick, 1964; Campbell, 1966; Seward, 1990; Dorobek, 1991; Pratt, 1998; Thomas, 2011; Grader and others, 2014). Stromatoporoids are some of the most distinctive and common fossils in the Jefferson Formation (figs. 14A–14E) and some genera, *Amphipora* and *Stromatopora* in particular, can form meter-size, low-relief build ups locally. Other fossils in the Jefferson Formation include corals (e.g., *Macgeea* sp.), bryozoans, bivalves (e.g., *Grammysia*, *Paleoneilo*), fish fragments, ostracodes, calcispheres, crinoids, calcareous algae, stromatolites (thrombolites), brachiopods, and conodonts (fig. 11). The latter two are of great significance for biostratigraphy of the Frasnian and Famennian in Montana and will be revisited later in this report.

Mimicking its diverse fossil assemblage, the facies of the Jefferson Formation in western and southwestern Montana are described by a diverse set of adjectives, including thin- to thick-bedded (fig. 15), black, brown, dark gray and light gray, petroliferous, medium crystalline to coarsely crystalline, highly porous, granular-weathering, fetid, and intercalated yellow and pale pink (Berry, 1943; Sloss and Moritz, 1951; McMannis, 1955; Klepper and others, 1957; McMannis, 1962; Knopf, 1963; Sandberg, 1965; Mudge, 1972; Kellogg, 1992; Stickney and Vuke, 2017). These are all used to describe the mainly dolomite facies in the Jefferson, and dolostone, sandy dolomite, dolomitic limestone, limestone, carbonate algal buildups, carbonate breccias, and argillaceous limestone that are present locally (figs. 11, 14).

A similarly diverse set of adjectives was used when describing the lithology in northwest and central Montana (figs. 11, 15). There, the dolomite of this interval, with limestone, lime mudstone, and wackestones and packstones present locally, is described as light gray, gray, dark gray (weathers chocolate brown), very dark gray, thin-bedded, medium-bedded, locally massive, finely crystalline, moderately to coarsely crystalline, with fetid odor locally, seams of black chert, nodules of algae and cup corals, some anhydrite

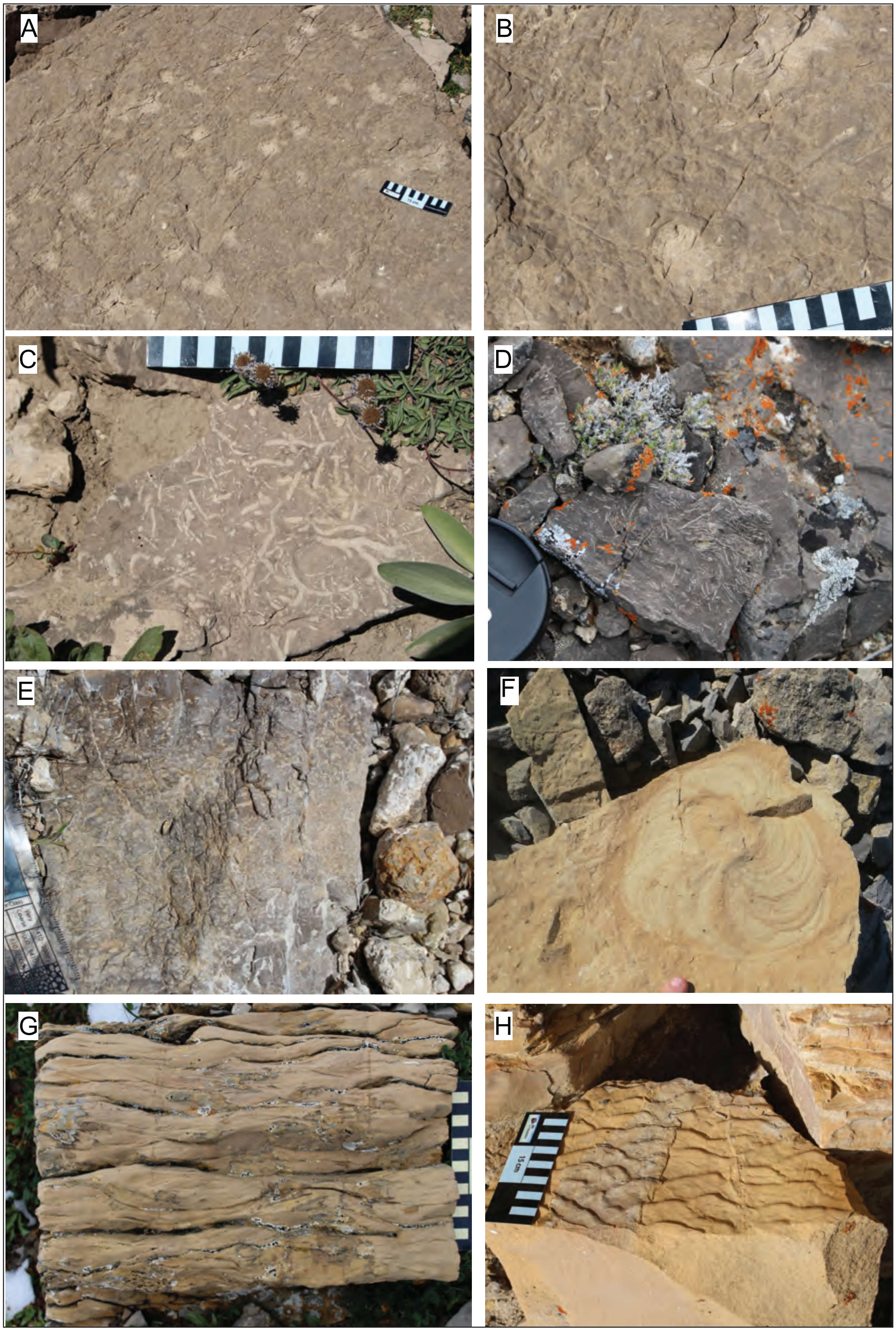

Figure 14. Photographs of typical Frasnian and Famennian facies in outcrop from the Jefferson Formation (A–E) and the Sappington Formation (F–H). Mottled, fossiliferous (stromatoporoids) dolomite from the Bridger Range (A) and close-up of the same sample (B). Fossiliferous (mainly tubular stromatoporoids) dolomite from the Bridger Range (C) and from near Ermont, SW Montana (D). Bioturbated, fossiliferous dolomite (E) from the Clark Canyon Reservoir area in SW Montana. Bioturbated (*Zoophycus*), dolomitic siltstone (F) is a common facies in the middle Sappington Formation (sample from southern Bridger Range, SW Montana). Ripple-laminated, dolomitic, very fine-grained sandstone from the Sappington Formation in the northern (G) and southern (H) Bridger Range, SW Montana, respectively.

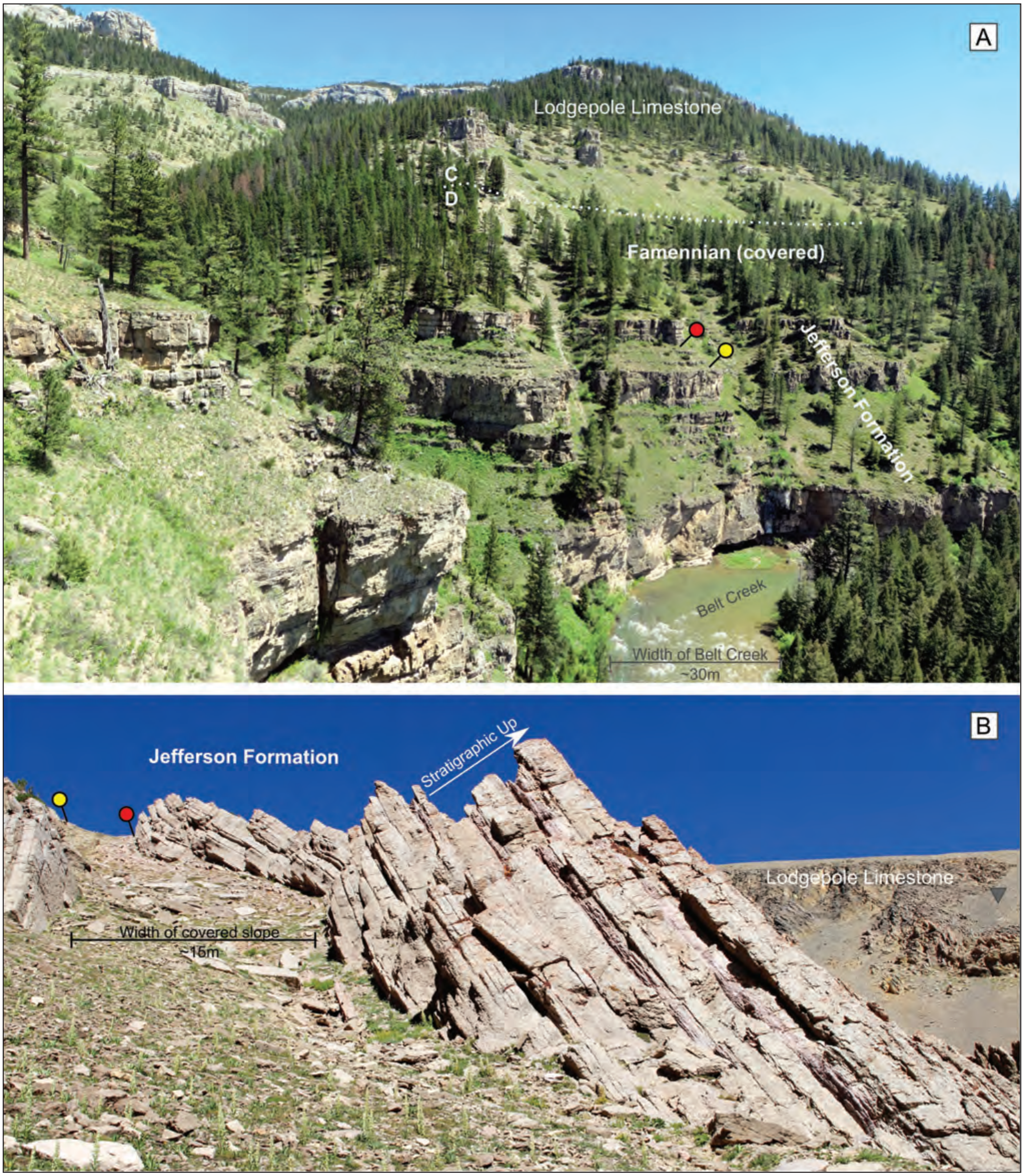

Figure 15. The Jefferson Formation in central (A) and southwest (B) Montana. (A) Cycles in the upper Jefferson Formation in the Little Belt Mountains exposed along Belt Creek, west of Monarch (see fig. 2 for location).The red pin marks the base of a depositional cycle near the top of the Jefferson Formation and, similar to the yellow pin, marks approximately the same lithostratigraphic position as shown in B. The Famennian strata (Three Forks and Sappington Formations) overlying the Jefferson Formation are less than ~30 m (~100 ft) thick in the Little Belt Mountains (Sandberg, 1965; Witkind, 1971; Sandberg and Poole, 1977) and, as shown in this photograph, are often covered by talus. The Devonian to Carboniferous boundary is covered and its approximate location marked by the white stippled line (D/C). The width of Belt Creek is approximately 30 m (~100 ft); the height of the first Jefferson cliff is approximately 15 m (50 ft). View is to the northeast. Photo source: author. (B) Photograph of depositional cycle hierarchy in the uppermost Jefferson Formation in the Bridger Range. The yellow and red pins mark approximately the same lithostratigraphic position as shown in photograph A (note the slope forming thin-bedded strata between the two pins in both locations), revealing a very similar bed thickness change (and cycle stacking) between these two locations. The red pin marks the approximate location of the base of a 3rd order cycle. The Lodgepole Limestone is exposed in a cliff in the background and is not unconformably overlying the Jefferson Formation as it appears in the photograph. The talus apron in the background (triangle) is the same apron marked by the triangle in figure 3 for reference. The width of the covered area is ~15 m (~50 ft). The view is to the northwest. Photo source: author.

or evaporite-solution breccia, and *Amphipora* facies and stromatoporoid biostromes (Sloss and Laird, 1947; Campbell, 1966; Witkind, 1971; Seward, 1990; Seward and Dyman, 1990; Kissling, 1996; Pratt, 1998; Porter and Wilde, 2001).

In the subsurface of the Williston Basin, the Jefferson Group is composed of largely brownish gray, finely crystalline dolomite, with lesser beds of gray to brownish gray microcrystalline limestone, yellowish to brownish gray, fine-grained argillaceous limestone and dolomitic limestone, white to gray anhydrite (fig. 16)—anhydrite is more common in the Birdbear (Nisku) Formation of the upper Jefferson Group—interbedded with thinner beds of greenish and yellowish gray dolomitic shale, very fine-grained siltstone, sandy argillaceous dolomite, and *Amphipora* facies and stromatoporoid biostromes (Sandberg and Hammond, 1958; Kissling, 1996).

Latest Devonian

The Three Forks Formation in western Montana (Peale, 1893; Sloss and Laird, 1947; Sandberg, 1962a, 1965), the Potlatch Anhydrite/Member in the Sweetgrass Arch area (Perry, 1928), the Three Forks Formation and the Torquay Formation in the Williston Basin (Christopher, 1961; Sandberg, 1965), and the Palliser Formation in northwestern Montana and western Alberta (Seward and Dyman, 1990; Savoy, 1992) all record earliest Famennian strata in Montana (figs. 5, 6).

In southwest Montana the Sappington Formation unconformably overlies the Three Forks Formation along a sharp contact between a prominent limestone bed in the upper Three Forks, and the dark lower shale of the Sappington Formation (fig. 3). Although the Sappington Formation was originally included with the Three Forks Formation at Logan Gulch (figs. 4, 12; Peale, 1893), it was recognized as a separate stratigraphic unit by Berry (1943), who also established a separate type section for the Sappington Formation. Although some subsequent workers regarded the Sappington as a member of the Three Forks Formation (e.g., Haynes, 1916; Sloss and Laird, 1947; Sloss and Moritz, 1951; Sandberg, 1962b, 1963, 1965; Klapper, 1966; Sandberg and Klapper, 1967; Gutschick and Rodriguez, 1990), others acknowledged it as a separate and younger stratigraphic unit with formation rank (e.g., Holland, 1952; McMannis, 1955; Achauer, 1959; Gutschick and Perry, 1957, 1959; McMannis, 1962; Gutschick and others, 1962; Gutschick and Rodriguez, 1967; Rodriguez and Gutschick, 1967; Smith and Bustin, 2000; Phelps, 2015; di Pasquo and others, 2017). This review does not attempt to solve this "Sappington Problem" (Sandberg, 1965), but solely uses the latter stratigraphic rank classification to emphasize and acknowledge the recent rise of the Sappington equivalent strata in northeastern Montana (Bakken Formation; Nordquist, 1953), as well as north-central Montana (Exshaw Formation; Warren, 1937; Sandberg, 1966) as one of the most prolific petroleum systems in North America (Gaswirth and others, 2013).

Three Forks/Palliser Facies

The Three Forks Formation and its time-equivalent strata are absent in central Montana and thicken to almost 305 m (1,000 ft) in northwestern Montana (Palliser Formation; fig. 5; Sandberg and Mapel, 1967; Seward and Dyman, 1990; Vuke and others, 2007).

Near its type locality in southwest Montana, the Three Forks Formation is subdivided into a lower Logan Gulch Member and an upper Trident Member composed of heterolithic strata of limestones, siltstones, and shale (figs. 3, 6). The Logan Gulch Member at its type locality consists largely of a mosaic of yellowish gray, grayish red, greenish gray shale and siltstone, argillaceous limestone breccia and shale breccia, silty dolomite, and dolomitic brownish gray stromatolitic limestone; the Trident Member is chiefly composed of greenish gray, light olive gray, and yellowish gray calcareous to slightly calcareous fossiliferous shale with brachiopods, cephalopods, ostracodes, crinoids, bryozoans, bivalves, and a rich ammonoid fauna (Peale, 1893; Sloss and Laird, 1947; Renzetti, 1961; Robinson and Barnett, 1963; Sandberg, 1965; Korn and Titus, 2006; Vuke and others, 2014; Thomas, 2011; di Pasquo and others, 2017). This change in facies represents the shift from restricted conditions in the Logan Gulch Member to more normal marine conditions in the Trident Member.

The mixed clastic-carbonate facies in the Three Forks type section gives way to mainly clastic deposition to the northeast (fig. 5). In the Little Belt Mountains of central Montana, near the present-day pinch-out of the Three Forks, the Logan Gulch Member is composed of light gray to green thin-bedded to platy sandstone with minor light gray dolomite interbeds (Witkind, 1971). To the west, calcareous facies (light and medium-dark gray bioclastic limestone and limestone breccia) become more abundant (Berg, 1959; Thomas, 2011). This trend of increased carbonate content continues into northwest Montana where the

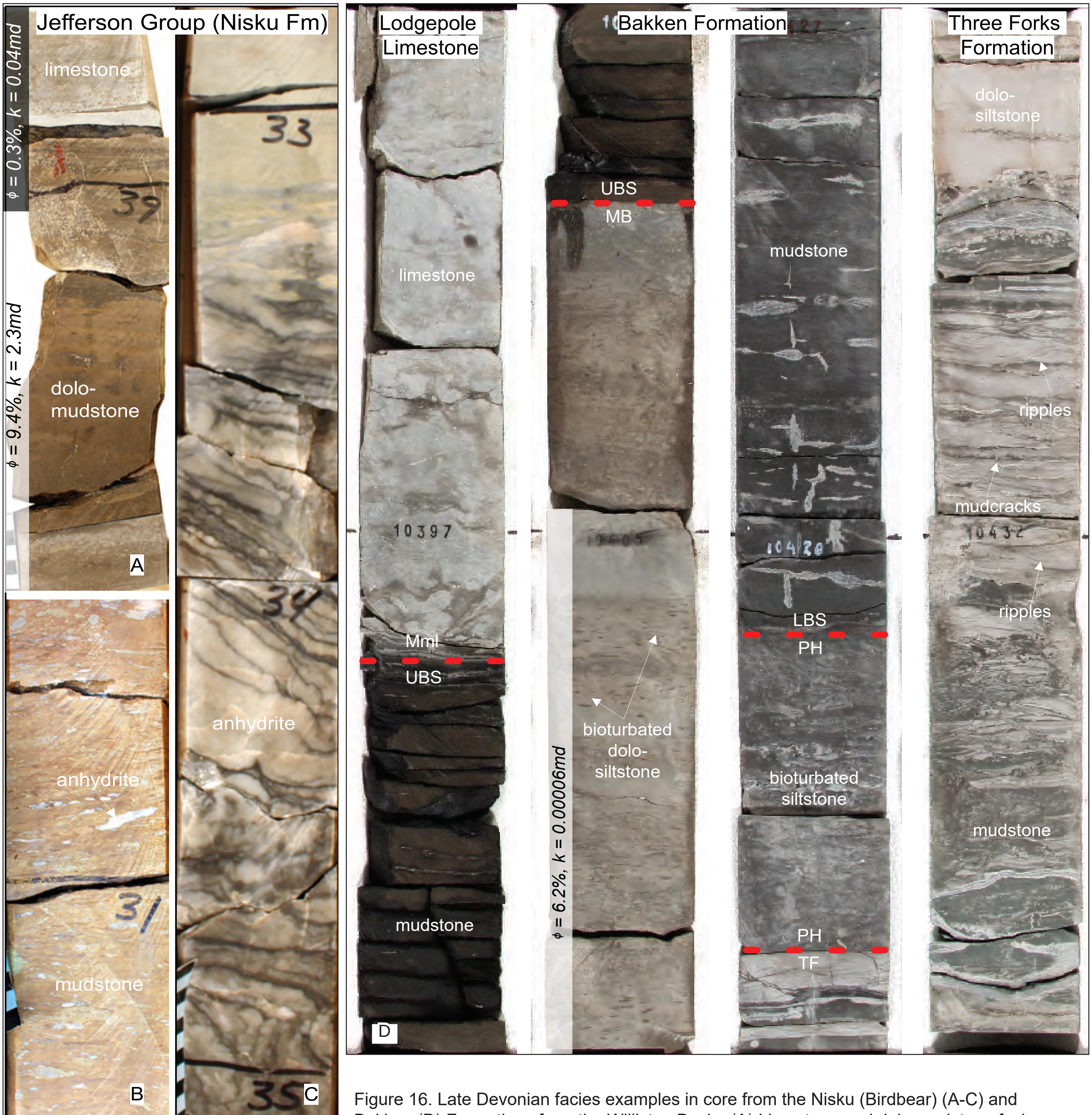

Figure 16. Late Devonian facies examples in core from the Nisku (Birdbear) (A-C) and Bakken (D) Formations from the Williston Basin. (A) Limestone and dolo-mudstone facies. The dolomites are porous with good permeability, whereas the limestones contain low porosity and permeability. (B) Mudstone facies with small anhydrite nodules. (C) Extensive anhydrite deposits. Facies B and C are more common towards the top of the formation.(D) The upper Bakken shale (UBS) consists of dark brownish gray to black mudstones and lighter colored siltstones. The middle member (MB) has abundant bioturbated dolo-siltstone and low porosity and permeability. The lower shale member (LBS) contains organic-rich mudstones and siltstones. A thin bioturbated siltstone interval is typical facies for the Pronghorn Member (PH) of the Bakken Formation. The Three Forks Formation consists of alternating dolo-siltstones and mud-stones with ripple lamination and mudcracks, evidence for fluctuating wetting and drying conditions during deposition. The red dashed lines mark lithostratigraphic formation boundaries, as well as sequence stratigraphic surfaces. In this case all surfaces are sequence boundaries of varying orders. The exact position of the D/C boundary in this core is unknown, but the UBS in other parts of the basin contains Tournaisian conodonts, whereas the LBS contains Famennian conodonts (e.g., Hogancamp and Pocknall, 2018). The lime-stones above the UBS are the basal limestones of the Mississippian Lodgepole Formation. Photos A–C from core Granley State 4-15X (API#: 2508521604), Roosevelt County, MT. Photos D from core Larson 11–26 (API# 2508322025), Richland County, MT (see fig. 2 for location). Scale bars (A–C) in centimeters. Core box height (D) is 2 ft. Photo source: author.

Palliser Formation (fig. 17) contains gray, greenish gray, olive, and brown partly dolomitized, micritized, and highly bioturbated peloidal/skeletal/fossiliferous lime mudstone and wackestone and packstones, and laminated, platy-weathering or bioturbated, nodular-bedded lime mudstone (Seward, 1990; Seward and Dyman, 1990; Savoy, 1992). Skeletal benthos (brachiopods, bryozoans, ostracods, phylloid algae, stromatoporoids) is, in general, more common in the upper Palliser Formation (Costigan Member) compared to the lower Palliser Formation (Morro Member; Seward and Dyman, 1990). This change in biota was interpreted to reflect more open marine conditions through time, similar to what is observed between the Logan Gulch and Trident Members of the Three Forks Formation in southwestern Montana.

The open marine facies of the Palliser Formation only occur in far northwest Montana (fig. 17). To the east along the international border, evaporites become increasingly abundant (fig. 5). In the Sweetgrass Arch area, massive anhydrite with nodular to chicken wire texture and dolomite interbeds are common (Schietinger, 2013). This is the area of the Potlatch Adams No. 1 well where Perry (1928) first recognized the presence of a thick unit of alternating shale, massive, pure, gray gypsum and anhydrite, mottled limestone, and porous dolomite that he named the Potlatch Anhydrite. The term Potlatch Anhydrite is still recognized locally in the Sweetgrass Arch area in northern Montana but is synonymous with and contemporary to the Logan Gulch Member. Farther to the east in the Little Rocky Mountains (fig. 2), evaporites decrease in abundance and are replaced by light gray and light green calcareous claystone, shale, and locally sandy siltstone (Porter and Wilde, 2001). Continuing east into the Williston Basin, the Three Forks Formation consists of olive, red-green, and gray shale that may be dolomitic, anhydritic, and silty (fig. 16). These facies are interbedded with tan, olive to red siltstone that may contain clay as well as very fine-grained sandstone grains, and are often dolomitic. Desiccation cracks are widespread in the mudstone facies and often filled by the dolomit-

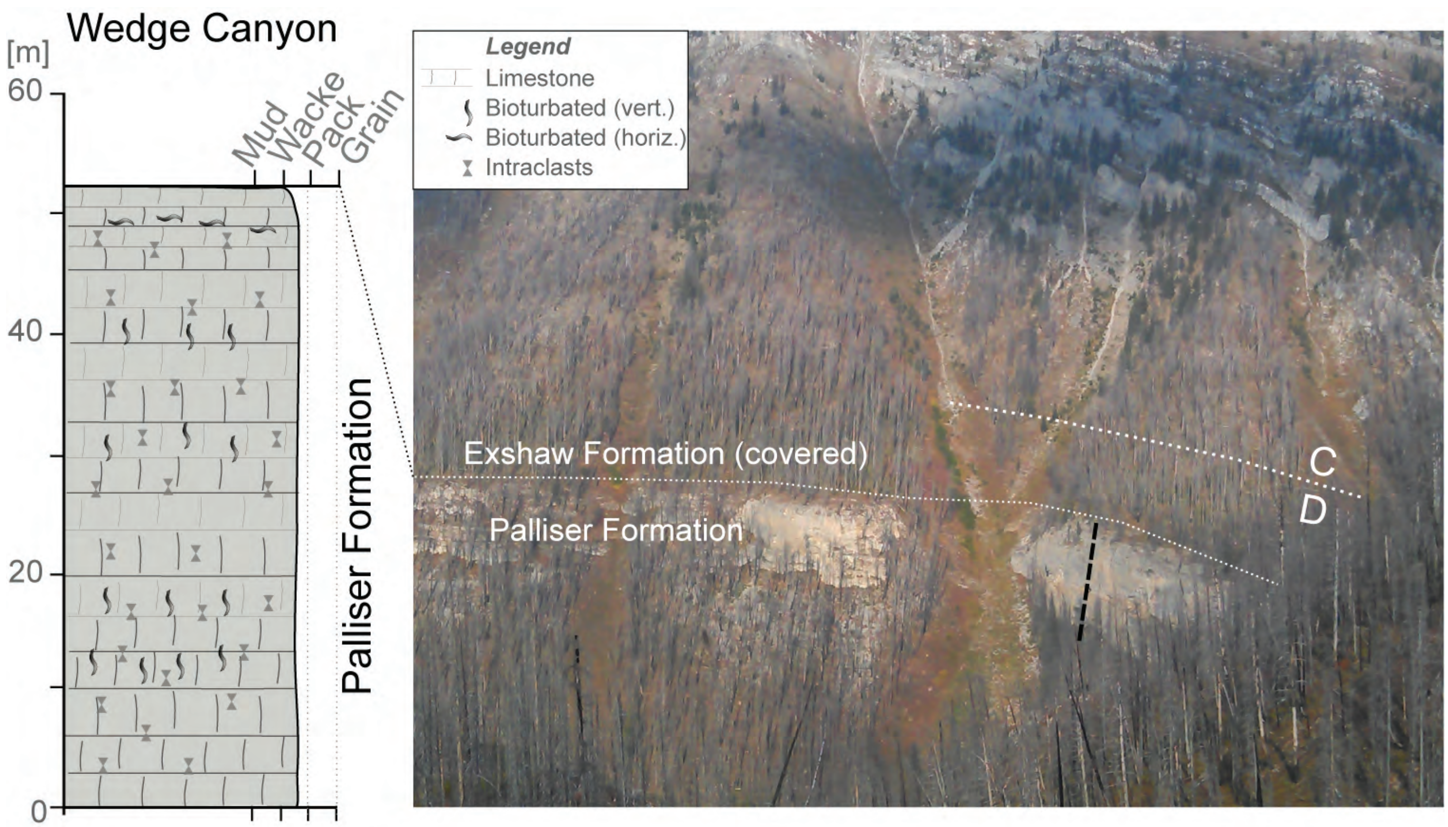

Figure 17. The Palliser Formation in the Whitefish Range in northwest Montana. The thickness of the Palliser Formation exceeds 200 m (650 ft) in an outcrop section near Eureka in the far northwest corner of Montana (Seward, 1990). Shown in this photograph are cliff-forming carbonates of the upper Palliser Formation (Costigan Member) in Wedge Canyon, just west of Glacier National Park (see fig. 2 for location). In this location the contact to the underlying Alexo Formation is covered, but where exposed the contact is conformable (Seward and Dyman, 1990). The contact to the overlying Exshaw Formation, the slope-forming unit just above the lower cliff, is disconformable. Seward (1990) measured approximately 52 m (~170 ft) of intraclast wacke-packstones of the uppermost Palliser Formation at this location. The approximate location of this measured section is shown by the black line (the exact location information is listed in Seward, 1990 and in Seward and Dyman, 1990). Measured section redrawn and modified from Seward, 1990. Photo source: author.

ic siltstone and sandstone, conglomerate and breccia, including dissolution breccia; and white anhydrite beds increase in abundance with depth (Sandberg and Hammond, 1958; Sandberg, 1961b; Dumonceaux, 1984; Webster, 1984; LeFever, 1991; Gantyno, 2010; LeFever and others, 2011; Franklin Dykes, 2014; Garcia-Fresca and Pinkston, 2016; Sonnenberg, 2017).

Sappington/Bakken/Exshaw

The lithologic succession of the latest Famennian strata is surprisingly consistent across the State (figs. 5, 6). In southwest Montana the Sappington Formation is composed of three distinct lithologic units, an upper and lower organic-rich black shale, and a middle dolomitic siltstone and sandstone (figs. 18, 19; e.g., Rodriguez and others, 2016; diPasquo and others, 2017, 2019; Phelps and others, 2018, and references therein). The Exshaw Formation in northwest Montana has very similar lithologies and is composed of an upper and lower organic-rich shale and a middle dolomitic siltstone member (e.g., Schietinger, 2013, and references therein). In its type well in the Williston Basin, the Bakken Formation also consists of ~6 m (20 ft) of slightly calcareous black shale (upper Bakken shale), ~18 m (60 ft) of calcareous, light gray to gray-brown, very fine-grained sandstone, interbedded with minor amounts of gray-brown, cryptocrystalline limestone (middle Bakken), and ~7.5 m (25 ft) of fissile black shale (lower Bakken shale; Nordquist, 1953). A similar lithology is present throughout most of the Williston Basin and into northeastern Montana, where the Bakken is composed of an upper and lower organic-rich black shale and a silty dolostone and dolomitic siltstone to sandstone middle member (figs. 16, 19). Recent studies have recognized a distinct mudstone and siltstone to sandstone interval below the Lower Bakken Shale as a fourth member, the Pronghorn Member (fig. 6; LeFever and others, 2011; Sonnenberg, 2017, and references therein). The Bakken Formation reaches its greatest thickness [~42 m (~140 ft)] in northwest North Dakota and thins towards the basin margins, including the western margins in Montana (Sonnenberg and others, 2017). In the Little Rocky Mountains of north-central Montana, the entire Bakken Formation is represented by a 1-ft-thick interval of black shale from the upper member, which unconformably overlies the Three Forks Formation and is overlain by the Lodgepole Limestone (Sando and Dutro, 1974).

In northwest and north-central Montana (fig. 5), in the Sweetgrass Arch area, the Exshaw Formation shows great lithologic similarity to the Bakken Formation (Sandberg, 1966). The Exshaw Formation in the

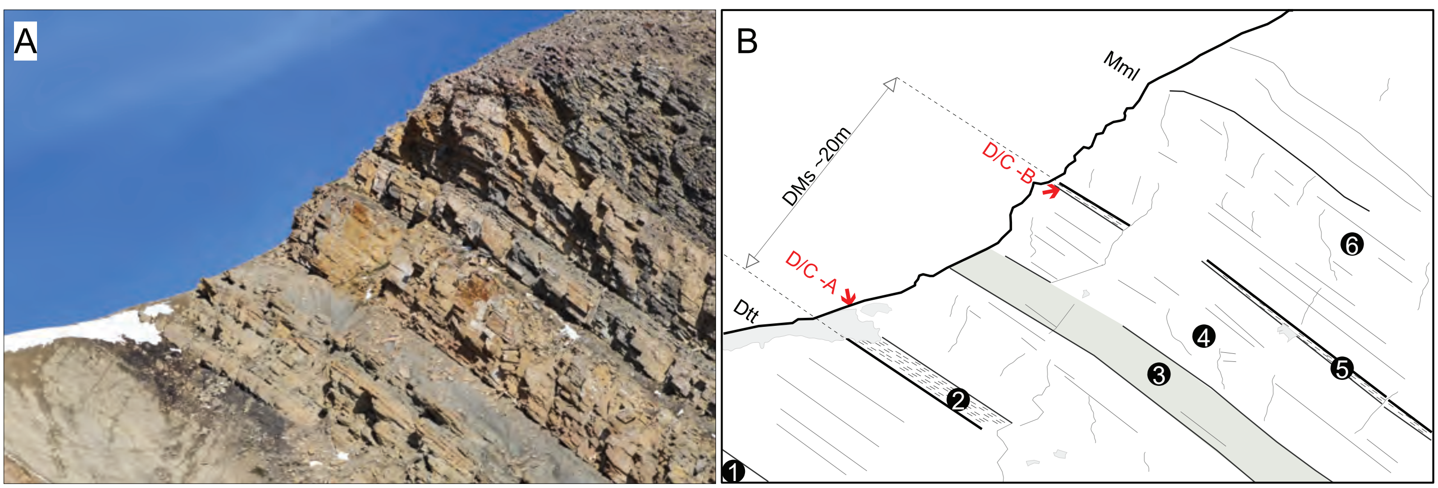

Figure 18. The latest Famennian and earliest Mississippian strata in the Bridger Range in southwest Montana. View is to the northwest; the entire Sappington Formation (DMs) is approximately 20 m (66 ft) thick. The top of the Logan Gulch Member of the Three Forks Formation—the ledge-forming carbonate—is visible in the lower left corner (1). The thin-bedded and laminar mudstones of the Trident Member (Dtt) are well exposed below the big snow patch. The lower Sappington shale (2) is organic-rich and easily recognizable by its black color. The middle Sappington contains two dolomitic siltstone and sandstone intervals that are separated by middle Sappington shale (3). The upper of the two dolomitic siltstone and sandstone intervals contains gradually dipping beds interpreted as clinoforms by Phelps and others (2018), (4). The upper Sappington shale (5) is another organic-rich shale that unconformably overlies the middle member and contains Tournaisian conodonts of the *Siphonodella crenulata* zone, placing the D/C boundary firmly in the Sappington Formation. Thin limestone beds of the Lodgepole Formation of the Mississippian Madison Group (Mml) are well exposed towards the top of the photograph. The color change from more orange-gray limestones to gray limestones (6) is a cycle boundary internal to the Lodgepole Limestone. D/C-A and D/C-B mark the location of the Devonian–Carboniferous (D/C) boundary in the Sappington Formation as interpreted through time (see text for references). Most recent studies place the D/C boundary at or near the D/C-B marker, but the higher uncertainty of biostratigraphic constraints in the middle Sappington result in a wide range of possible surfaces for this significant global boundary. For outcrop location see black dashed rectangle in figure 3. Photo source: author.

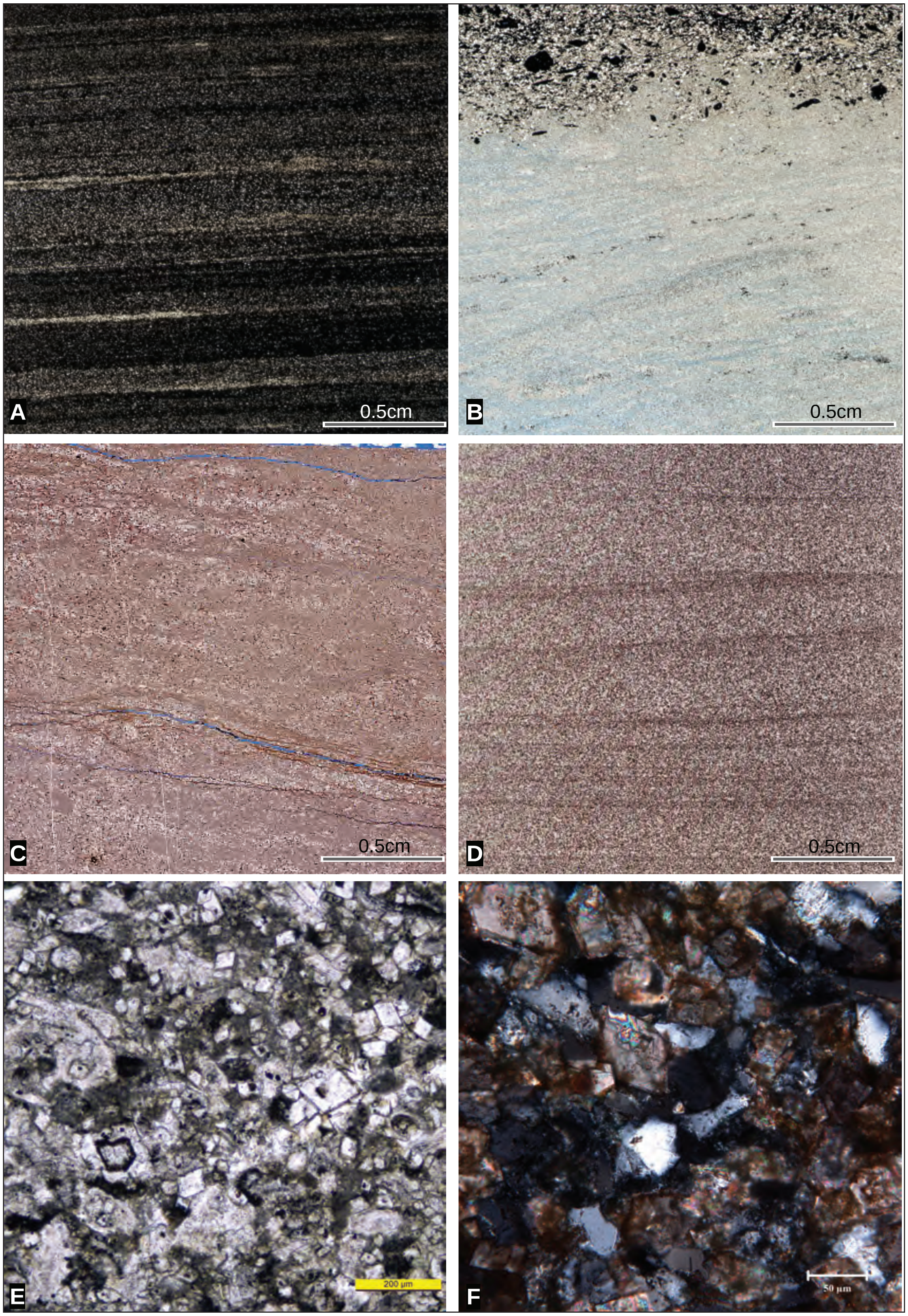

Figure 19. The Bakken and Sappington Formations in thin section from northeastern (A, B), south-central (C, E) and southwestern (D, F) Montana. The upper Bakken shale (A) in northeastern Montana (core Williams 1–4, API#: 2508321676, Richland County, MT) contains abundant siltstone laminae (lighter colored) between the organic-rich mudstones (darker color). The organic-rich facies in this sample contains ~11% TOC. In the same core, the contact between the middle Bakken dolomitic siltstones and the upper Bakken shale is marked by a lag deposit with abundant phosphatic grains (B). The abundance of dolomite in the middle member is not limited to the subsurface in northeastern Montana, but is also common in the middle member of the Sappington Formation in south-central and southwest Montana. This thin section scan (C) shows mottled facies in the Sappington from the Boulder River in Sweetgrass County (see fig. 2 for location). The plane polarized light photograph of the same sample (E) shows the abundance of (zoned) dolomite crystals in this facies. A similar composition and replacement history can be observed in the cross-laminated sample (D) from the Bridger Range in southwest Montana (see fig. 2 for location). Euhedral dolomites fill pores and replace detrital grains as is clearly visible in this cross-polarized light photomicrograph (F).

Sweetgrass Arch area is composed of an organic-rich lower shale, a middle siltstone, and an organic-rich upper shale. The thickness of the Exshaw in this area varies greatly (Schietinger, 2013). The lower shale thickness ranges from approximately 30 cm (1 ft) in the southern Toole County and northern Pondera County, to more than 6 m (20 ft) in northern Toole County, just south of the international border and west of the Sweetgrass Arch. The middle siltstone unit, which pinches out in southern Toole County, ranges to a thickness of more than 30 m (100 ft) in northern Toole County just south of the international border and west of the Sweetgrass Arch. The thickest accumulation of the upper shale is farther to the west and in far northeastern Glacier County, where it reaches just over 3 m (10 ft). The upper shale quickly thins to the south and east and is largely absent in Pondera and Teton Counties to the south and the Sweetgrass Arch to the east (Schietinger, 2013).

In southwestern Montana the Sappington Formation follows the same lithologic tripartite of latest Famennian strata observed to the north and east (figs. 6, 18). Around the type locality in Milligan Canyon, near Three Forks, the Sappington is ~15–30 m (50–100 ft) thick and is composed of organic-rich black shale, green-gray slightly calcareous shale and siltstone, and gray-orange to yellow-brown siltstone and very fine- to medium-grained sandstone—sandstones and siltstones are locally dolomitic and/or calcareous (e.g., Berry, 1943; Sloss and Laird, 1947; Holland, 1952; McMannis, 1955; Sandberg, 1965; Klapper, 1966; Adiguzel and others, 2012; Nagase and others, 2014; Rodriguez and others, 2016; diPasquo and others, 2017, 2019; Phelps and others, 2018). South of the Three Forks area, the Sappington Formation thins to less than 12 m (40 ft; McMannis and Chadwick, 1964; Rodriguez and others, 2016), and eventually pinches out onto the Beartooth Shelf. The thickness changes in this region are accompanied by distinct facies changes. The thicker, more complete Sappington Formation sections contain largely dolomitic siltstone, ripple-laminated siltstone, bioturbated siltstone, and less common very fine- to fine-grained sandstone in the middle member (figs. 14, 19), accompanied by organic-rich mudstones in the lower and upper shales. Thinner sections in outcrop contain a higher abundance of trough cross-bedded sandstone with coated grains (ooids) also present in the middle Sappington, and a higher abundance of siltstone in addition to mudstones as part of the lower and particularly the upper shale (e.g., Gutschick and Rodriguez, 1990; Rodriguez and others, 2016; Phelps and others, 2018).

To the north of the Sappington type area, the thickness of the Sappington Formation remains relatively constant at ~15–21 m (50–70 ft) in the northern Boulder Batholith area and the northern Big Belt Mountains (fig. 2; Knopf, 1963; Sandberg, 1965), before dramatically thinning to the northeast, in the Little Belt Mountains (fig. 15) and the Big Snowy Mountains. There, only parts of the middle and upper Sappington are preserved, and the Sappington Formation is mainly composed of pale red to light brown thin-bedded siltstone, and ranges in thickness from ~4 m to 7.5 m (~13 ft to 25 ft; Witkind, 1971; Sandberg and Poole, 1977; Nagase and others, 2014).

AGE CONTROL

Age control of Devonian rocks in Montana is largely based on biostratigraphy, including conodonts, brachiopods, palynomorphs, ammonites, and vertebrates, and more recently from detrital zircons. In outcrop, in the western half of the State, most of the Devonian strata is Late Devonian (Frasnian and Famennian). In the east, in the subsurface of the Williston Basin and eastern Montana, Middle and Late Devonian strata are preserved, resulting in the most complete Devonian sections being recorded from core in the subsurface.

Early and Middle Devonian

The Beartooth Butte Formation represents the oldest Devonian unit in Montana (figs. 5, 6). Early studies assigned deposition of this unit to the Pragian (Sandberg and Mapel, 1967; Mallory and Hennerman, 1972), but more recent work on vertebrate fossils and spores has shown that the Beartooth Butte Formation is not one contemporaneous deposit. Instead, it contains Emsian fossil assemblages in the type section at Beartooth Butte in Wyoming, and Pragian fossils in the Cottonwood Canyon locality in the Big Horn Mountains in Wyoming, just south of the Montana border (Tanner, 1983; Elliot and Ilyes, 1996; Elliot and Johnson, 1997). A similar asynchronous deposition has been reported from the Beartooth Butte Formation in east-central Idaho, with vertebrates of early Early Devonian (Lochkovian) age described in the Lost River Range, and vertebrates of late Early Devonian (Emsian) age described from the Lemhi Range close to the Montana border (Grader and Dehler, 1999).

The Middle Devonian strata in Montana are poorly constrained, owing to deposition in nearshore and stressed environments that precluded widespread deposition of sediment with normal marine biota. Most information about the biostratigraphy comes from sparse microfauna (conodonts) and macrofauna (brachiopods and stromatoporoids) from limited well-bore data in the Williston Basin, or from outcrop data in Canada. For example, circumstantial stratigraphic evidence seems to suggest an Eifelian deposition of the Ashern Formation in the Williston Basin of North Dakota (figs. 5, 6; Lobdell, 1984), but the depositional age of the Ashern was interpreted to range anywhere from Silurian to Devonian (e.g., McGehee, 1949; Baillie, 1951; Norris and others, 1982; Lobdell, 1984; Rosenthal, 1987). The Winnipegosis Formation is commonly assigned an upper Eifelian (*Polygnathus ensensis* conodont zone) to early Givetian (*Pol. ensensis* to *Pol. varcus* zone) age in the Williston Basin (Jones, 1965; Ehrets and Kissling, 1987, Stearn, 2001), although a time-transgressive deposition that reflects the ongoing Kaskaskia I transgression onto the Montana craton is inferred (Perrin, 1982, 1987). The Prairie Evaporite conformably overlies the Winnipegosis Formation and is most likely middle Givetian (*Pol. varcus* conodont zone; Ehrets and Kissling, 1987; Stearn, 2001). A contact of debated conformity—as discussed previously, the presence (significance) of an unconformity is debated in literature (e.g., Sandberg and Mapel, 1967; Ehrets and Kissling, 1987)—separates the Dawson Bay Formation from the Prairie Formation in the Williston Basin (figs. 5, 6), but a middle Givetian (upper *Pol. varcus* zone) to late Givetian deposition is still most likely (Ehrets and Kissling, 1987; Stearn, 2001).

The ages of the Maywood and Souris River Formations are constrained by sparse brachiopod, fish, and conodont faunas, and from stromatoporoid biostratigraphy in western Canada. Early studies suggested Cambrian and Silurian deposition for the Maywood Formation (Emmons and Calkins, 1913; Dorf and Lochman, 1940). However, the presence of *Bothriolepis, Charophytes*, fish scales, brachiopods of the *Allanaria allani* zone, conodonts of the *Polygnathus asymetricus* zone (*falsiovalis* zone) and *Palmatolepis disparilis* zone, and stromatoporoids of the species *Actinostroma expansum* place the upper Maywood and upper Souris River Formations in the late Givetian and lower Frasnian (figs. 5, 6; Lochman, 1950; Wilson, 1956; Klepper and others, 1957; Freeman and others, 1958; Sandberg and Hammond, 1958; Sandberg and McMannis, 1964; Meyers, 1971; Mudge, 1972; Stearn, 2001). Biostratigraphic age control on the lower parts of the Souris River and Maywood Formations is less abundant, but a late Middle Devonian (late Givetian) deposition is commonly inferred, with time-transgressive deposition starting in western Montana and becoming younger eastward onto the Montana craton (Meyers, 1971).

Late Devonian

The ages of Late Devonian strata are well constrained biostratigraphically. Conodonts of Frasnian age are abundant in the Jefferson Formation of western Montana, comfortably placing it in the *Palmatolepis transitans* zone to the *Pa. rhenana* zone (figs. 5, 6; Grader and others, 2016). Frasnian brachiopods are also common throughout the Jefferson Formation/Group and include *Cyrtospirifer* sp. (common in the Duperow Formation), *Tenticospirifer, Atrypa multicostellata kottlowski*, *Atrypa* sp. (common in the Birdbear Formation/Member), *Eleutherokomma* cf., *Eleutherokomma reidfordi*, and *Productella* (e.g., Cooper and others, 1942; Kottlowski, 1949; McMannis, 1962; Sandberg and Mapel, 1967; Mudge, 1972; Hogancamp and Pocknall, 2018).

Lower and Middle Famennian

A similarly rich fauna of ammonites, brachiopods, and conodonts is described from outcrops of the Three Forks Formation. Although mainly absent in the Logan Gulch Member, they are common in the Trident Member. Conodonts from the type section of the Trident Member are middle Famennian *Palmatolepis marginifera* to *Pa. trachytera* zones (Klapper, 1966; Sandberg and Klapper, 1967; Sandberg and others, 1972; Johnston and others, 2010). The same authors assign the Logan Gulch Member to early Famennian *Pa. triangularis* to *Pa. marginifera* zones. A similar depositional age is reported from the Palliser Formation on the Alberta Shelf in northwest Montana. There, the lower Palliser Formation (Morro Member) contains conodonts from the early Famennian *Pa. crepida* to *Pa. marginifera* zones (Savoy, 1992; Savoy and others, 1999), suggesting deposition synchronous with the Logan Gulch Member (figs. 5, 6). In addition to conodonts are diverse ammonite and brachiopod faunas, including ammonites of the *Platyclymenia annulata* zone in the Trident Member, ammonites of the *Cheiloceras* ammonite zone found in the Logan Gulch Member in southwest Montana, and the brachiopod *Pugnoides minutus* found in the Sun River Canyon area (fig. 2). Each of these biozones constrains the Three Forks Formation to the Famennian (Raymond, 1907, 1909; Schindewolf, 1934; Miller, 1938; Benson, 1966; Mudge 1972; Korn and Titus, 2006). More recent work on the palynomorph assemblage of the Trident Member also supports a middle Famennian depositional age for these rocks (di Pasquo and others, 2017).

As well correlated as the Three Forks Formation is in the western part of Montana, correlating across the Central Montana Uplift is notoriously difficult and has led to more uncertainty in the correlation of facies and members from western Montana into the Williston Basin (Sandberg and Hammond, 1958; Sandberg, 1965). Recent studies indicate that the Three Forks Formation in the Williston Basin is largely correlative to the Logan Gulch Member in southwest Montana (Sandberg and others, 1988; Hartel and others, 2012; Sonnenberg, 2017), but there is less certainty on the correlation of the Trident Member into the Williston Basin (figs. 5, 6). Although some studies have shown that the Trident Member pinches out to the east and has no equivalent strata in the Williston Basin (Sandberg and others, 1988; Hartel and others, 2012; Sonnenberg, 2017), other studies have suggested that the Trident Member is correlative to the Big Valley Formation in the northern Williston Basin in Saskatchewan (Johnston and Chatterton, 2001; Grader, 2014). Yet other studies have suggested that the Big Valley Formation of Saskatchewan is equivalent to the Pronghorn Member of the Bakken Formation in North Dakota and Montana (LeFever and others, 2011; Gaswirth and Marra, 2015), and the Three Forks Formation in the southern and eastern Williston Basin is equivalent to the Torquay Formation in Saskatchewan and part of the upper Palliser Formation in Alberta (*Pa. marginifera* and *Pa. trachytera* conodont zones; Smith and Bustin, 2000). The stratigraphic framework of the Three Forks Formation becomes even more complicated in the subsurface, because of the petroleum industry's habit of dividing the Three Forks into lithostratigraphically defined "benches" that often lack biostratigraphic control or sequence stratigraphic relevance. Depending on the author and operator, the Three Forks in the subsurface is subdivided into between three and six informal stratigraphic units (e.g., Webster, 1984; Gantyno, 2010; Sonnenberg, 2017). A good summary of the Three Forks stratigraphic nomenclature currently in use in the subsurface was provided by Gaswirth and Marra (2015).

Late and Latest Famennian

Despite the great lithologic similarity among the Sappington, Bakken, and Exshaw Formations, the precise ages of these formations are less certain, and the question as to whether these three formations are contemporaneous or are merely coincidental lithologic equivalents has been at the center of the debate since the first discovery of these rocks (figs. 5, 6). Brachiopods were used early on to assign an age to the Sappington Formation, and conodonts and palynomorphs were added more recently to put depositional age constraints to the Sappington, Bakken, and Exshaw Formations in Montana and adjacent areas (Haynes, 1916; Banta, 1951; Holland, 1952; Morgridge, 1954; McMannis, 1955; Achauer, 1957, 1959; Harker and McLaren, 1958; Gutschick and others, 1962; Sandberg, 1965, 1976, 1979; Klapper, 1966; Sandberg and Klapper, 1967; Sandberg and others, 1972, 1988, 2002; Sandberg and Ziegler, 1979; Huber, 1983; Hayes, 1985; Thrasher, 1987; Karma, 1991; Playford and McGregor, 1993; Savoy and Harris, 1993; Drees and Johnston, 1996; Johnston and others, 2010; Warren, 2015; Rodriguez and others, 2016; di Pasquo and others, 2017, 2019; Hogancamp and Pocknall, 2018).

In northwest Montana conodonts from the Exshaw Formation span from the latest Famennian (*Palmatolepis expansa* zone) to the early Tournaisian (*Siphonodella crenulata* zone; Seward, 1990; Savoy, 1992; Savoy and Harris, 1993; Savoy and others, 1999; Johnston and others, 2010; Schietinger, 2013), without any large depositional gaps reported based on conodont zonation (fig. 5).

Early studies of the Bakken and Sappington assigned the entire Sappington Formation to the Tournaisian (Holland, 1952), similar to some early studies of the biostratigraphy of the Bakken Formation in the Williston Basin (Nordquist, 1953) and some more recent studies in southeastern Montana (Macke, 1993). However, most biostratigraphic studies assign the depositional age of these formations to the latest Famennian and early Tournaisian. For example, Achauer (1959) reported conodonts of the *S. sandbergi* to *S. quadruplicata* conodont zones (conodont zones sensu Ziegler and Sandberg, 1984) from the upper middle Sappington siltstone and conodonts of the *S. sandbergi* to *S. quadruplicata* conodont zones from the upper shale. In another early study, a conodont assemblage at the base of the lower shale indicates deposition during the *Pa. expansa* zone of Famennian (Klapper, 1966; Sandberg and Klapper, 1967). Recent composite conodont biostratigraphy charts have placed the lower Sappington shale either entirely within the *Pa. expansa* zone (di Pasquo and others, 2017, 2019; Phelps and others, 2018), or extending downward into the *Pa. postera* zone (Rodriguez and others, 2016). Palynology studies place the middle Sappington member into the *S. praesulcata* zone (Warren and others, 2014; Warren, 2015; diPasquo, 2017), and the upper shale into the Tournaisian (diPasquo and others, 2012). Because of its close temporal relationship (Tournasian) to the overlying Lodgepole Limestone, the upper Sappington shale also can be correlated to the Cottonwood Canyon Member of the Lodgepole Limestone (Sandberg and Poole, 1967; Grader and others, 2016).

Conodonts recovered from the Bakken Formation in the Williston Basin place the lower Bakken shale in the *Pa. trachytera* zone (Hogancamp and Pocknall, 2018), synchronous with the Trident Member of the Three Forks Formation in western Montana (figs. 5, 6). This latest proposed Bakken Formation depositional age is significantly older than suggested by previous researchers, who placed the lower Bakken shale in the *Pa. gracilis expansa* conodont zone (Hayes, 1985; Huber, 1983; Thrasher, 1987). These studies also report partial conodonts of the *S. crenulata* zones from the upper shale, suggesting deposition contemporaneous with the Lodgepole Limestone. However, recovered conodonts of the *S. quadruplicata* conodont zone from the same interval in the Williston Basin suggest an asynchronous deposition of the Bakken and the Lodgepole Limestone (Hogancamp and Pocknall, 2018). This interesting debate about the depositional age of this latest Devonian and earliest Mississippian strata in Montana will certainly continue for decades as new age dating methods are developed and applied to these rocks.

PALEOGEOGRAPHY—TECTONIC SETTING AND EUSTACY

Montana during the Early Devonian—A Place Exposed

A widespread regression at the end of the Ordovician marked the end of the Tippecanoe cratonic sequence and left Montana widely exposed during much of Silurian and Early Devonian deposition (fig. 5). During the onset of the Kaskaskia transgression, heterolithic sedimentary layers of the Beartooth Butte Formation (fig. 7) infilled valleys, channels, crevices, and sinkholes in this barren landscape. Most of the Beartooth Butte was deposited in estuaries and rivers—fluvial deposits include channel fill and floodplain deposits, as well as paleosols (Dorf, 1934; Sandberg 1961a; Sandberg and Mapel, 1967)—with a transition from more freshwater conditions in central Montana to more brackish conditions southward into northern Wyoming (fig. 7; Fiorillo, 2000). The Early Devonian [Pragian (Lochkovian) to Emsian] was a time of tectonic quiescence in Montana but a time of frequent eustatic changes (fig. 6; Johnson and others, 1985; Dorobek, 1991; Haq and Schutter, 2008). As a result, the facies changes in the Beartooth Butte Formation—both spatially and through time—likely reflect these eustatic changes at multiple frequencies. The depositional history of the Beartooth Butte Formation remains enigmatic, and new methods and a fresh set of geologic eyes likely will continue to provide surprises about the true nature of these rarest of Devonian rocks in Montana.

The Middle Devonian—A Period of Episodic and Local Flooding of Montana

In the Middle Devonian (late Eifelian to Givetian), marine waters of the Kaskaskia transgression inundated Montana (fig. 8). The first area affected in the State was the Elk Point Basin in northeastern Montana, which formerly connected open marine waters in the Northwest Territories of Canada to the Williston Basin (fig. 8). This initial flooding of the intracratonic basin is reflected by the transgressive lag deposits at the base of the Ashern Formation. A continuing stepwise flooding of the basin and alternation between restricted and normal marine conditions are recorded by evaporites and red beds interbedded with carbonates bearing cephalopods, gastropods, and brachiopods. The gradual change upward from mainly reddish deposits and evaporites near the base of the Ashern Formation to more green deposits and fewer evaporites was interpreted to reflect gradual deepening during the initial transgression of the Kaskaskia megasequence (fig. 6; Sloss, 1963; Lobdell, 1984; Johnson and others, 1985; Megathan, 1987; Haq and Schutter, 2008).

The overall eustatically driven, time-transgressive deepening in the Williston Basin continued into the Givetian, and the thin basal siliciclastics and mixed siliciclastic-carbonate of the Ashern Formation were largely replaced by the carbonates of the Winnipegosis Formation (Baillie, 1953; Johnson and others, 1985; Haq and Schutter, 2008). The Winnipegosis carbonates in Montana were initially deposited on a marginal ramp that, over time, evolved into a broad shelf that deepened to the north into southern Saskatchewan and Manitoba (Jones, 1965; Perrin, 1987; Ehrets and Kissling, 1987). In Montana, on the slopes and shelves of the eastward (present day) deepening basin (fig. 8), the Winnipegosis carbonate facies is largely reflective of restricted lagoonal, shallow shoal, interbioherm, bioherm, deep shoal, restricted shallow water, and tidal flat environments (fig. 9; Sandberg and Hammond, 1958; Sandberg, 1961b; Kinard and Cronoble, 1969; Perrin, 1982, 1987; Oster, 2016).

Eustatic rise from the ongoing Kaskaskia I transgression played an important role in the initial Winnipegosis Formation deposition, but the control on sedimentation is less certain during the latest stages of the Winnipegosis deposition. A basinward shift of facies belts and progressively shallower facies, including intertidal to supratidal carbonates and anhydrites, and the formation of more isolated basins, suggest the end of transgressive conditions during the latest Winnipegosis deposition (Perrin, 1982, 1987; Ehrets and Kissling, 1987). This overall shoaling, caused by regional uplift, decreased subsidence, eustatic sea-level stillstand or fall, or a combination of all of these factors, resulted in the formation of shallow restricted basins on the craton. These restricted conditions culminated in widespread evaporite deposition (Prairie Formation; fig. 9) across the Elk Point and Williston Basins of northeastern Montana (Sandberg and Hammond, 1958; Holter, 1969; LeFever and LeFever, 2005; Luo and others, 2017).

The relative contributions of eustasy vs. local subsidence/uplift associated with preexisting structural lineaments (figs. 6, 8) in the shoaling and closure of the basin remain unclear because observed thickness changes in the Prairie Formation could reflect either depositional variations or salt dissolution (Baillie, 1953; LeFever and LeFever, 2005). One key to better understanding this record of basin isolation and evaporite deposition is a thin sequence of red to green non-fossiliferous dolomite and calcareous shale, the so-called Second red bed (Oglesby, 1988; LeFever and LeFever, 2005) that tops the evaporites (fig. 6). This unit could be a residual product of salt solution and weathering, or it could be the result of increased clastic input related to transgression and deposition of the overlying Dawson Bay Formation (Baillie, 1953; Holter, 1969; Oglesby, 1988).

Following widespread deposition of evaporites of the Prairie Formation, the Kaskaskia transgression resumed during late Middle Devonian (Givetian) time, but with a geologic record still only preserved in the eastern part of the State. In the Williston Basin, this phase of sea-level rise resulted in the deposition of the Dawson Bay Formation on a stable, low-relief shelf that extended into northeastern Montana. Ultimately, this northwestern extension of the Williston Basin reconnected with marine water of the Elk Point Basin and mixed clastic-carbonate deposits of the basal Dawson Bay Formation accumulated (figs. 6, 9). The transition from the argillaceous deposits in the lower Dawson Bay to the carbonates and evaporite deposits near the top of the Dawson Bay Formation suggests eventual restriction of the Williston Basin, which Maughan (1989) interpreted to reflect another brief regression near the end of the Middle Devonian (fig. 6).

The Late Devonian Stepwise Drowning

In striking contrast to the restriction of Middle Devonian strata almost entirely to the subsurface of northeastern Montana, Late Devonian strata are widespread across the State (fig. 5). The more regional preservation of Late Devonian rock across Montana resulted from continued transgression of marine water onto the North American craton during the Kaskaskia I supersequence (fig. 6), resulting in the establishment of an epicontinental sea with two transgressive wedges, one from the east and one from the west, that coalesced in central and north-central Montana.

In the east, in the Williston Basin, the Souris River Formation was deposited on an eastward-dipping (present-day geography) basin margin, overlying the Middle Devonian Dawson Bay Formation (figs. 6, 8, 10). The contact between the two formations is conformable towards the center of the basin, but unconformable towards the margins and into north-central Montana (figs. 5, 6; e.g., Mallory and Hennerman, 1972; Ehrets and Kissling, 1987).

In western Montana, the transgressive Kaskaskia seas inundated the undulating Cambrian to Silurian subcrop topography (figs. 5, 8) that was formed during the long-lived exposure of the broad cratonic shelf. Significant localized thickness changes of the Maywood Formation are the result of this transgression over a largely erosional topography (fig. 10; McMannis, 1962; Sandberg and McMannis, 1964; Grant, 1965; Meyers, 1971, 1980; Thomas, 2011).

The same complex basal bathymetry/topography resulted in the formation of bays, estuaries, and tidal flats with brackish or even freshwater conditions (Sandberg and McMannis, 1964; Meyers, 1980). The presence of basal conglomerate beds in some areas in western Montana (Meyers, 1971; Thomas, 2011), as well as eastern central Montana (Sandberg and Hamond, 1958; Sandberg, 1961b), also suggest at least short-lived fluvial conditions that filled some of the incised valleys during early transgression. This continental interpretation for the basal part of the Maywood and Souris River Formations is further supported by the presence of exposure surfaces with weakly developed soils in places (Lochman, 1950; Hanson, 1952). Recent detrital zircon studies from the Maywood strata in western Montana reflect a complex provenance of major tectonic provinces across the North American continent (Hendrix and others, 2016), indicative of the initial reworking of a long-exposed passive margin shelf (Ingersoll, 1990; Ingersoll and others, 1993).

Gradual deepening as the Kaskaskia transgression continued resulted in mainly intertidal to subtidal carbonate of the middle Maywood Formation. Supratidal dolostone caps the Maywood in western Montana, and anhydrite occurs in the upper Souris River Formation in eastern Montana, suggesting a return to shallower (Maywood) and/or restricted conditions (Souris River).

The Montana-wide Kaskaskia I transgression (fig. 6) that started slowly during Middle Devonian (late Givetian) reached its climax during the Late Devonian (Frasnian; Johnson and others, 1985; Sandberg and others, 2002).

The Jefferson Formation carbonate was deposited during this maximum Kaskaskia I transgression, when nearly all of Montana was inundated by a shallow sea (fig. 6; Johnson and others, 1985; Sandberg and others, 2002). The Jefferson Formation was deposited on a shallow carbonate platform and progressively onlapped westward and eastward towards the platform apex in central Montana from present-day Idaho and the Williston Basin, respectively (Webb and Wilhelm, 1983; Dorobek, 1991; Grader and others, 2016). As a result of this enduring asynchronous transgression, the oldest Jefferson deposits in Idaho are Middle Devonian (Eifelian; Johnson and others, 1985; Dorobek, 1991), whereas in central Montana deposition of the Jefferson did not occur until the Late Devonian (Frasnian; fig. 5).

Superimposed on the gradual changes in the depositional system—and as a result gradual facies changes—associated with the long-lived Kaskaskia I transgression onto the craton are high-frequency depositional cycles (fig. 15). This high-frequency cyclic deposition started with the initial Kaskaskia transgression, became the norm during deposition of the Souris River and Maywood Formations (fig. 6), and is best documented in the Jefferson Formation/Group (e.g., Peterson, 1984; Dorobek, 1991; Grader and others, 2014). In Montana these cycles are composed of largely shallow marine limestones that grade westward into deeper marine facies (Dorobek, 1991). Individual cycles in the Jefferson shoal upward and vary in thickness from less than 1 m to tens of meters and are interpreted as 3rd order cycles (e.g., Grader and others, 2014). This record of cyclicity and varying thicknesses is well visible in outcrops across the State. Figure 15 shows examples of the cyclic deposition in the upper Jefferson in the Bridger Range and in the Little Belt Mountains. Individual cycles within the Jefferson thin upward, from massive to thick-bedded closer to the base of cycles to thick- to thin-bedded near their tops. This is clearly visible in the example from Belt Creek (fig. 15A). These depositional cycles are less than a meter to tens of meters thick, not just at these two locations, but throughout southwestern Montana (Dorobek, 1991). Cycle stacking shows an overall upward thinning of cycles and beds. Note the massive beds near the base of the exposure (massive cliff-forming bed) directly above the cut bank of Belt Creek is approximately 15 m (~50 ft) in height and more dominant medium- and thin-bedded strata near the top of the Jefferson (fig. 15A).

The lowest order cycles in the Jefferson Formation are interpreted as 3rd order cycles that can be recognized regionally (fig. 15B; e.g., Dorobek, 1991; Grader and others, 2014). These 3rd order cycles are composed of stacked higher order cycles that can be interpreted from bed thickness changes and stacking.

Although these 3rd order (and higher order) cycles can be recognized in many locations, correlating individual cycles over longer distances can be challenging. The complexity and uncertainty when trying to correlate and interpret Devonian stratigraphy across Montana is injected, in part, by the presence of local paleogeographic highs that greatly influenced thickness and facies of the Jefferson carbonate cycles on the gradually dipping carbonate ramps (e.g., Sandberg and Poole, 1977; Webb and Wilhelm, 1983; Ehrest and Kissling, 1983; Peterson, 1986; Dorobek and Smith, 1989; Dorobek, 1991; Grader and others, 2014, 2016). These local uplifts are the result of the onset in tectonism (Antler Orogeny) associated with the active continental margin development that occurred west of Montana during Middle and Late Devonian time (fig. 6; Beranek and others, 2016). Figure 20 shows one possible correlation of Jefferson 3rd order cycles between two outcrop locations, Logan and Mill Creek (see fig. 2 for location), that is based on the changes in cycle thickness trends (Dorobek, 1991). Individual cycles are characterized by thicker carbonates (dolomitic packstones and wackestones) at their base. These carbonates are bioturbated and contain normal marine skeletal debris and biota. The tops of cycles are marked by thinner beds composed of finer-grained facies, often stromatolite–thrombolite–cryptalgal laminatites (microbialites), indicating more stressed conditions and also contain clear signs of subaerial exposure (dissolution collapse breccias). The similar varying thickness of these cycles is convincing; however, the controlling factors of individual cycles—eustatic sea-level changes, tectonic changes, and local basin conditions among others—remain nebulous and need to be considered in the global context (see also discussion on global events in section 5 of this report).

The Latest Devonian (Famennian)— The Kaskaskia II Supersequence

During the Late Devonian, the North American continent was largely inundated by the floodwaters of the late Devonian Kaskaskia transgression (fig. 21A). However, during the Frasnian, the stable passive continental margin of western North America that influenced deposition throughout most of the Devonian became a convergent margin (figs. 6, 21). Associated with the convergent margin is the Antler Orogeny, a contractile mountain-building event that began in the Late Devonian and is commonly attributed to the accretion of the island arc to the western continental margin (Johnson, 1971; Poole, 1973; Johnson and Pendergast, 1981; Sandberg and others, 1986; Dorobek, 1991; Jansma and Speed, 1993; Smith and others, 1993; Beranek and others, 2016). East of the allochthonous Antler highland, in present-day Nevada and Idaho, the Antler Foreland Basin developed as a result of crustal flexure. In Montana the Antler Fore-

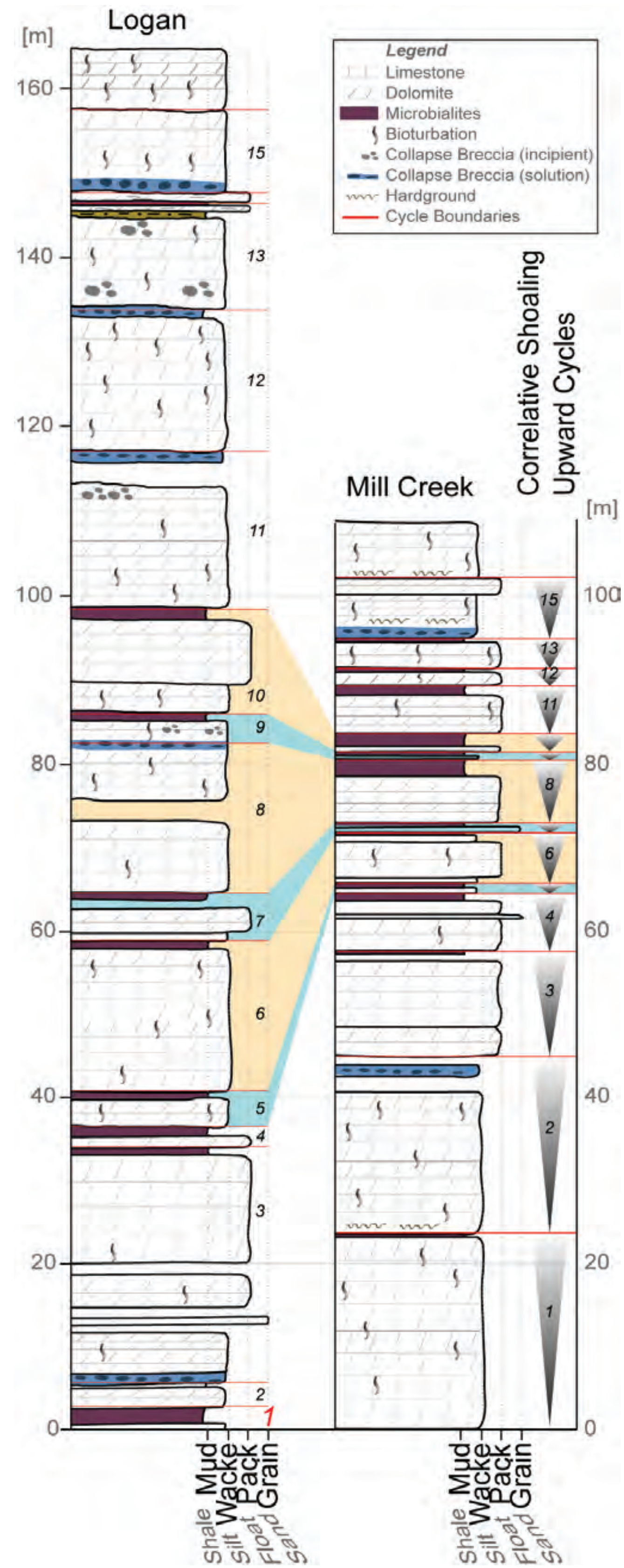

Figure 20. Correlation of depositional cycles in the Jefferson Formation between Logan and Mill Creek (see fig. 2 for location), Correlative intervals are highlighted in yellow and blue. Also compare the measured section shown in this figure to the outcrop photographs of Jefferson cycles, from different locations, in figure 15. Sections and correlations redrawn and modified from Dorobek (1991) and incorporating lithologic information from Sandberg (1965).

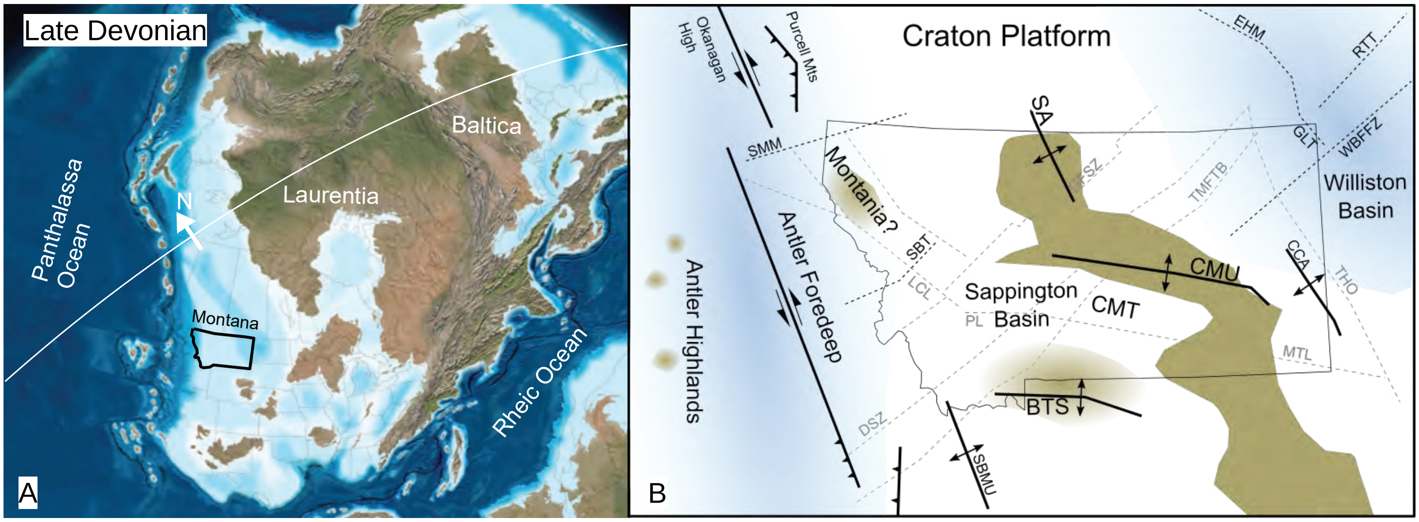

Figure 21. Late Devonian paleogeography of North America (A) and latest Devonian paleogeography and structural lineaments in Montana and adjacent areas (B). Paleozoic lineaments (black): BTS, Beartooth Shelf (Yellowstone uplift); CCA, Cedar Creek Anticline; CMT, Central Montana Trough; CMU, Central Montana Uplift; EHM, Elbow–Hummingbird Monocline (Hummingbird trough); GLT, Goose Lake trends; RTT, Rocanville–Torquay trend; SA, Sweetgrass Arch (Bearpaw Anticline); SBMU, South Beaverhead Mountain Uplift (Lemhi Arch); SBT, Scapegoat–Bannatyne trend; SMM, St. Mary–Muvie transform; WBFFZ, Weldon–Brockton–Froid fault zone. Precambrian lineaments (gray): DSZ, Dillon Shear Zone; GFSZ, Great Falls Shear Zone; LCL, Lewis and Clark line; MTL, Mesoproterozoic Montana–Tennessee line; PL, Perry line; THO, Trans-Hudson Orogen; TMFTB, Trans-Montana fold and thrust belt. Maps compiled and modified from McMannis (1965), Sandberg and Mapel (1967), Mallory and Hannerman (1972), Gerhard and others (1990), Kent and Christopher (1994), Sims and others (2004), Fischer and others (2005), Beranek and others (2016), and Blakey (2016).

land Basin geometry was complicated by diachronous crustal loading, and the presence of crustal lineaments that were oblique or perpendicular to the axis of the Antler Foreland Basin (fig. 21B; Peterson, 1986; Dorobek, 1991; Trexler and others, 2004). As such, Famennian deposition was even more strongly influenced by local uplifts and basins than its Frasnian predecessor. Uplifts that appear to have influenced deposition during the latest Devonian, like the Beartooth Shelf (BTS) and the Central Montana Uplift (CMU), are aligned with (parallel to) known basement structures like the Perry line (PL) and the Mesoproterozoic Montana–Tennessee line (MTL), and are oblique (nearly perpendicular) to the main Antler orogenic front (fig. 21B). This apparent reactivation of structural lineaments and a subsequent control on facies distribution extends all the way into the Williston Basin,d where known late Devonian depositional centers like the Hummingbird trough (EHM) and facies divides like the Weldon–Brockton–Froid fault zone (WBFFZ) follow a similar orientation as prominent basement structures like the Trans-Hudson Orogen (THO), and the Trans-Montana fold and thrust belt (TMFTB). The presence of these inherited structures resulted in the complex depositional history in isolated basins as described for the latest Devonian and is a diversion from the largely eustatic control on deposition that was present throughout most of the passive margin phase of Middle and early Late Devonian deposition.

The Three Forks Formation and its time-equivalent strata, the oldest Famennian strata in Montana, were deposited following the latest Frasnian maximum transgression (fig. 6). This regressive phase at the onset of the Kaskasia II sequence (T-R cycle IIe of Johnson and others, 1985), and the paleogeographic position near the equator (fig. 21), resulted in widespread deposition of marine, restricted marine, and evaporite deposits of the lower Three Forks Formation (Logan Gulch Member) and correlative strata (Vuke and others, 2007). In contrast, the Trident Member (upper Three Forks) in southwest Montana is generally interpreted as reflecting open marine conditions (Sandberg, 1965), although analysis of ammonite assemblages, palynology, and low-diversity bryozoan colonies indicate that stressed marine conditions occurred at least episodically during the upper Three Forks deposition (Prezbindowski and Anstey, 1978; Korn and Titus, 2006; di Pasco and others, 2017).

In northwestern Montana the deposits of the Palliser Formation—contemporary to the Logan Gulch Member (figs. 5, 17)—have been largely interpreted as subtidal and peritidal facies deposited on a westward-dipping carbonate ramp (Savoy, 1992; Savoy and others, 1999), representing the overall deepening of the craton into the Antler Foreland Basin. A local unconformity at the base of the Palliser Formation in northwest Montana (Seward, 1990), and the presence of stem fragments (arborescent lycopsid) in the Morro Member (lower Palliser Formation) just north of the international border, suggest episodic exposure of parts of the shelf during the Famennian (Hartel and others, 2012, 2014; Pratt and van Heerde, 2017). The spatial and temporal extent of the subaerially exposed area are both speculative (Ross and Stephenson, 1989; Johnston and others, 2010; Hartel and others, 2012, 2014; Grader and others, 2014; di Pasco and others, 2017), but might coincide with the position of Deiss' (1941) Cambrian Montania landmass (fig. 21).

East of the inferred local Montania uplift—the location and presence of the Montania landmass is poorly constrained in literature but its position might align with the Lewis and Clark line (LCL) and transform faults along the Antler orogenic front (fig. 21)—and west of the Sweetgrass Arch, a contemporaneous restricted evaporitic basin developed containing the widespread evaporite deposits of the Potlatch Anhydrite (Logan Gulch Member equivalent; McMannis, 1962; Sandberg and others, 1988; Schietinger, 2013). To the south, the restricted evaporitic facies wanes towards the Logan Gulch type section near Three Forks and transitions into offshore evaporitic facies (these evaporites are preserved as solution breccias; e.g., di Pasquo and others, 2017). Farther to the south towards the Beartooth Shelf, the Logan Gulch Member facies are interpreted as a near-shore evaporitic facies (McMannis, 1962; Robinson and Barnett, 1963). This shoaling trend to the south is oblique to the axis of the Antler Foreland Basin, but a similar shoaling trend also occurs to the east onto the Central Montana Uplift.

East of the Central Montana Uplift, in the Williston Basin, the depositional environment interpretations of the Three Forks facies are as variable as the facies themselves. A cyclic depositional style is inferred by many, although little consensus has emerged as to the depositional environment of the Three Forks Formation in the Williston Basin. Interpretations have included a low-relief shelf (Christopher, 1961), a tide-dominated environment with subtidal sand flats and intertidal mudflats (Bottjer and others, 2011), a melange of depositional conditions including intertidal to subtidal environments, terrestrial paleosols, sabkha and subaerial gravity flows (Egenhoff and others, 2011), a shallow shelf and mudflat complex with eustacy-controlled fluctuating hypersaline and freshwater conditions (Franklin Dykes, 2014; Franklin and Sarg, 2017), a playa lake depositional model (Garcia-Fresca and Pinkston, 2016), and high frequency (4th order) sequences superimposed on a 3rd order deepening upward cycle, with the depositional environment changing from predominantly supratidal to intertidal (Gantyano, 2010; Sonnenberg, 2017). The diversity of interpretations of the Three Forks depositional setting in the Williston Basin is not surprising, given that the only datasets available are from the subsurface and the absence of good biostratigraphy increases the uncertainty of whether these different researchers were describing contemporaneous strata.

It is entirely possible that a complex basin geometry might be responsible for a much more spatially fragmented deposition in the Williston Basin than has been documented to date. A simple thermal subsidence model is most often evoked for the intracratonic Williston Basin (Crowley and others, 1985; Xie and Heller, 2009), but local stratigraphic thinning associated with crustal lineaments (fig. 21) suggests that at least some areas were decoupled from a homogeneous thermal subsidence (Sandberg, 1964; Kent, 1987; LeFever and Crashell, 1991). Although a direct relationship between structural partitioning within the Williston Basin and far-field tectonics related to Antler flexure has yet to be documented, linking the facies heterogeneity of the Three Forks across Montana to tectonic forcing or, conversely, to eustatic forcing will be an exciting area of future research.

The continued eastward migration of the Antler Orogen during the latest Famennian exerted a flexural load on the craton and reactivated crustal lineaments (fig. 21) that are not aligned with the axis of the Antler Foreland Basin (Peterson, 1986; Dorobek, 1991; Trexler and others, 2004). This crustal loading produced asynchronous and localized subsidence in Montana, resulting in complex depositional conditions in the Bakken, Exshaw, and Sappington Basins, respectively, that reflect the changing eustatic and tectonic controls, spatially and through time (figs. 5, 6).

In the southwest, in the Sappington Basin (figs. 2, 21), the basal contact between the lower Sappington shale and the underlying Trident Member of the Three Forks Formation is interpreted by many researchers as a prominent sequence boundary (figs. 6, 18) based on the missing conodonts of the *Pa. postera* conodont zone (e.g., Johnston and others, 2010; di Pasquo and others, 2017, 2019). Numerous workers have inferred that the shale members of the Sappington Formation represent an offshore marine depositional environment with anoxic and dysoxic bottom waters (Gutschick and Rodriguez, 1979; Grader and Doughty, 2011; Adiguzel, 2012; Nagase and others, 2014; Myrow and others, 2015; Rodriguez and others, 2016; di Pasquo and others, 2017, 2019; Phelps and others, 2018; Browne and others, 2020). More controversial are depositional environment interpretations of the middle member siltstone, which include shallow marine to tidal flat and deltaic (Grader and Doughty, 2011; Adiguzel, 2012), lagoonal (di Pasco, 2017), and offshore transition zone to upper shoreface environment (Nagase and others, 2014; Rodriguez and others, 2016; Phelps and others, 2018). The different interpretations are, in part, the results of the datasets on hand for these studies. For example, Phelps and others (2018) based their interpretation mainly on facies and stratigraphic architecture relationships from outcrop, di Pasquo and others (2017) based their interpretation on facies and palynomorphs, and Cole and others (2015) and Myrow and others (2015) based their interpretation on facies and isotopes. An even more complicated picture emerges from recent detrital zircon work by Romenus and others (2020) on the middle member of the Sappington Formation. This work suggests a provenance associated with an easterly (Appalachian) source. The mechanism and timing of sediment dispersal across the Williston Basin, across the Central Montana Uplift, and into the Sappington Basin, and how they are linked to eustacy (fig. 6), reactivation of basement structures (fig. 21), and/or flexural loading during the Antler Orogeny are yet unknown and have become an exciting twist to the already complex latest Devonian history in Montana.

Like the Sappington Formation, the depositional setting of the Bakken Formation in the Williston Basin has been the subject of decades-long debate that includes both organic-rich shale members and the middle Bakken siltstone and sandstone member. Unfortunately, to date, no reliable biostratigraphic or chronostratigraphic data have been recovered from sandstone and siltstone of the middle Bakken member (Hart and Hofmann, 2020). This absence of biostratigraphic or other chronostratigraphic evidence in the middle Bakken member makes it impossible to accurately correlate depositional surfaces among the Bakken, Exshaw, and Sappington Formations.

Within the Williston Basin, the organic-rich lower and upper Bakken shales are interpreted to be marine (e.g., Hart and Steen, 2015; Sonnenberg and others, 2017; Hogancamp and Pocknall, 2018), although redox conditions are debated. Recent work on elemental data suggests that deposition occurred below storm wave base in a basin with largely anoxic to euxinic bottom (pore) waters (Scott and others, 2017). However, other workers have suggested at least episodic oxygenation of bottom waters as evident by bioturbation and winnowed lags in the shales (Egenhoff and Fishman, 2013; Hogancamp and Pocknall, 2018). The middle Bakken member deposition is most often interpreted as reflecting a shallow-marine setting (e.g., Smith and Bustin, 1998; Angulo and Buatois, 2012; Sonnenberg and others, 2017; Hogancamp and Pocknall, 2018; Hart and Hofmann, 2020), but restricted lagoonal conditions might have occurred locally (Hogancamp and Pocknall, 2018). Although mappable (erosion) surfaces are observed within the middle Bakken member, these are poorly constrained geometrically and chronologically across the basin (e.g., Hart and Hofmann, 2020).

The many interpretations might be the result of local depositional controls imposed by local subsidence differences across the Williston Basin that might be controlled by basement structures (fig. 21), as was documented for the Pronghorn Member of the Bakken Formation in North Dakota (LeFever and others, 2011; Hofmann and others, 2017). But it also opens the door for multi-proxy and collaborative research approaches that might help shed light on the depositional environment and processes in these basins while helping to better constrain the structural and depositional history in Montana during the latest Devonian and earliest Mississippian.

THE MONTANA ROCK RECORD OF GLOBAL DEVONIAN EVENTS

Lithostratigraphic, biostratigraphic, and sequence stratigraphic correlations of Devonian strata across the State are limited by patchy reliable absolute age control, uncertainty of biostratigraphy, post-depositional erosion, and limited rock access (in particular in the subsurface of eastern Montana). Although these limitations are an obstacle to deciphering the controls on deposition within Montana, they are also a significant impediment to linking the Montana rock record to events that not just affected the rock record locally in Montana, but also influenced deposition along the craton margin, the entire North American craton, and throughout the global record. With an understanding and acceptance of these limitations, the following attempts to connect the Devonian of Montana to some important global Devonian events as well as to indicate some current uncertainties.

D/C Boundary

The placement of the Devonian to Carboniferous (D/C) boundary (fig. 6) is of great interest to many scientists around the globe, because it correlates with global environmental changes (e.g., Hangenberg crisis) and faunal mass extinctions that had extinction rates between 45% (e.g., Sepkoski, 1996) and 70% (Kaiser and others, 2020). The uncertainty surrounding the causes of mass extinctions at the end of the Devonian requires not just a detailed understanding of the environmental changes that occurred in a basin, but also when they occurred and whether they are contemporaneous or developed diachronously across the globe. The latter depends on a detailed biostratigraphic framework, but across the globe, the D/C boundary is often within conodont-free siliciclastic intervals, or the first occurrence of relevant index fossils is uncertain (e.g., Kaiser and others, 2020).

Similarly, in Montana the D/C boundary is not associated with a distinct formation boundary (figs. 6, 18), but instead is commonly placed within the siliciclastic-dominated latest Famennian to earliest Tournaisian Sappington/Bakken/Exshaw Formations (e.g., Banta, 1951; Achauer, 1959; Gutschick and others, 1962; Hogancamp and Pocknall, 2018; di Pasquo and others, 2019). Most biostratigraphic control in these formations comes from the upper and lower marine shales. In contrast, biostratigraphic control of the middle siltier and sandier unit is poor. Therefore, it is not surprising that over time, stratigraphic placement of the D/C boundary has varied. For example, in the Sappington Formation in southwest Montana (fig. 18), the D/C boundary has been placed at the base of the Middle Sappington Member (fig. 2; Gutschick and Rodriguez, 1967), at or near the top of the Middle Sappington (Sandberg and Klapper, 1967; Rodriguez and others, 2016; diPasquo and others, 2017, 2019), and in the middle of the Middle Sappington (Klapper, 1966; MacQueen and Sandberg, 1970; Sandberg, 1979). Similarly, the D/C boundary was originally placed at the base of the Bakken Formation in the Williston Basin (Nordquist, 1953; Kume, 1963), before being moved to the middle of the middle Bakken member (Thrasher, 1987; Holland and others, 1987), and most recently was placed near the base of the middle Bakken (Hogenkamp and Pocknall, 2018).

This uncertain placement of the D/C boundary in Montana, and similarly anywhere in the world, has far-reaching implications when interpreting strata and processes affecting the Devonian strata regionally or globally (figs. 5,6).

5.2 Events, Ice Ages, Orogenies, Eustacy

The Devonian Period is well known as the Age of Fishes and the rapid spreading of land-plants, but it was also a time of major mass extinction and oceanic anoxic (black shale) events (e.g., Raup and Sepkoski, 1982; Buggisch, 1991; Sepkoski, 1996; Becker and others, 2016; Kaiser and others, 2016, 2020), glaciations (e.g., Caputo and others, 2008; Isaacson and others, 2008; Brezinski and others, 2010), and major tectonism (e.g., McKerrow and others, 2000; Miall, 2019), most of which either controlled frequent sea-level changes in the Devonian (e.g., Johnson and others, 1985; Johnson and Sandberg, 1988; Sandberg and others, 2002; Brett and others, 2020), or were associated with (resulted from) sea-level changes (fig. 6). However, these often contemporaneous events can make it difficult to clearly segregate the multiple controlling mechanisms from the rock record.

The Devonian rocks in Montana are no exception, and as significant as some of these events might have been, a signal in the rock record is not always distinct and easy to observe. For example, the onset of the Antler Orogeny and the change in structural regime from a passive margin to an active margin (Dorobek, 1991; Beranek and others, 2016) is recorded in the Frasnian and Famennian strata in Montana, namely the Jefferson Formation/Group and younger strata. However, the Late Devonian Gondwana glaciation (e.g., Streel and others, 2000; Caputo and others, 2008; Isaacson and others, 2008) and a drop in sea surface temperature coincided with the change in structural style (fig. 6). Which of these events had a greater effect on the shift from a carbonate-dominated Devonian world (Jefferson and older formations) to an environment more strongly influenced by siliciclastic deposition (figs. 5, 6; Three Forks, Sappington Formations) remains unresolved, as does how this dual control is recorded in the cyclic deposition of Late Devonian rocks in Montana (figs. 15, 20; e.g., Dorobek and Smith, 1989; Dorobek, 1991; Grader and others, 2014, 2016).

This meshing (or amplification) of multiple signals in the rock record of Montana applies not just to tectonism or glaciation, but also to geologically short-term events such as the two most significant Devonian first order bioevents (mass extinctions) and oceanic anoxic events (fig. 6; Becker and others, 2016), the Kellwasser crisis at the end of the Frasnian (e.g., Buggisch, 1991; Girard and Renaud, 2007; Carmichael and others, 2019), and the Hangenberg crisis at the end of the Famennian (e.g., Kaiser and others, 2016; Zhang and others, 2020).

In Montana the Kellwasser anoxic event(s), and the Frasnian–Famennian mass extinction event, coincide with a basinward shift in facies (regression, recorded stepwise at the base of the Birdbear Member of the Jefferson Formation and the base of the Three Forks Formation, respectively (fig. 6). Although often linked to changes in climate (e.g., Joachimski and Buggisch, 2002; Joachimski and others, 2009; Huang and others, 2018), a combination of controlling factors more likely explains this anoxic event (Carmichael and others, 2019). How and where exactly the Kellwasser crisis is expressed in the different paleoenvironments recorded in these Frasnian rocks across Montana, and how the plurality and interdependency of factors are going to be decoupled, are fascinating topics for future research.

Similar important research is needed to unravel the signal of other Late Devonian crises, particularly the Hangenberg event (fig. 6). As briefly mentioned previously, the Hangenberg crisis is another major global environmental change and mass extinction at the D/C boundary (e.g., Brett and others, 2020). This event has been observed and correlated around the world (e.g., Becker and others, 2016), and global anoxia has emerged as a likely cause based on the presence of organic-rich shales in many sections (e.g., Kaiser and others, 2016; Zhang and others, 2020). As discussed earlier in this review, recent studies of Montana stratigraphy have suggested that despite their great lithologic similarity, contemporaneous deposition of the Bakken, Sappington, and Exshaw Formations is questionable (fig. 5). Therefore, the signal of the Hangenberg crisis likely has a unique signature in the rock record in different parts of the State. Within the current constraints of biostratigraphy, in southwestern Montana the Hangenberg event is correlated to the middle Sappington shale bracketed by siltstones and sandstones (e.g., diPasco and others, 2019), whereas in the Williston Basin in northeastern Montana, the Hangenberg event is placed in the lower black shale of the Bakken Formation (fig. 6; e.g., Hogancamp and Pocknall, 2018). The discrepancy in lithologic response to a global event likewise extends to less significant bioevents (Becker and others, 2016), such as the Dasberg and Annulata events (fig. 6). Although in southwestern Montana the Dasberg event is correlated to the lower black shale of the Sappington Formation, and the Annulata event is suggested to correlate with the Three Forks Formation (diPasco and others, 2019), in northeastern Montana the same two events are interpreted to both be recorded in the lower black shale of the Bakken Formation (Hogancamp and Pocknall, 2018). These interpretations of global events in the Montana rock record highlight some interesting questions. Why do these global events that have a very similar lithologic response in other parts of the world have such inconsistent lithologic responses in different parts of Montana? Are these global events truly contemporaneous or are they periods of changes that are asynchronous across the globe and driven by local changes? What is the uncertainty of the biostratigraphic and chronostratigraphic constraint of Devonian rocks in Montana, and are the lithologic responses of these events truly different?

The original goal of this review was to summarize over a century of excellent Devonian geologic research in Montana and how it relates to large-scale controlling factors, namely eustacy and tectonics. Over the past century we have learned a lot about the Devonian in Montana, but many questions pertaining to the lithostratigraphic, biostratigraphic, and sequence stratigraphic record of Devonian rocks in Montana remain. In particular the manifestation of global events in the Devonian of Montana is a fascinating, yet still poorly understood topic. Therefore, continued study of the Devonian of Montana is not just critical to advance the understanding of the geologic evolution of Montana during this exciting and complex time in the Earth's history, but to help further advance our understanding of the multiple controlling mechanisms that shaped the landscape and biologic evolution more than 360 million years ago. With plenty about Devonian rocks in Montana yet to discover, "this completes, so far as I am aware, the list of those who have done Devonian geological work in Montana to date" (quote modified from Peale, 1893).

ACKNOWLEDGMENTS

Reviews by Mike Pope and Bruce Hart, as well as comments and edits by Susan Vuke and Marc Hendrix on earlier versions, helped to improve the manuscript and are greatly appreciated. Final edits by editor Susan Barth are greatly appreciated. Susan Smith and Meghan Rorick are thanked for their work on illustrations. Marc Hendrix provided access to several thin section samples. The many geologists studying the Devonian of Montana over the past century are thanked for the treasure trove of information that made this review possible.

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