

PASSIVE TRANSGRESSION: REMARKABLE PRESERVATION AND SPATIAL DISTRIBUTION OF UPPERMOST DEVONIAN (FAMENNIAN) MARGINAL AND NEARSHORE MARINE FACIES AND FAUNA OF WESTERN LAURENTIA

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ABSTRACT: The stratigraphic record of shallow-water shoreline paleoenvironments is characterized by significant facies heterogeneity and laterally discontinuous stratal geometries. In contrast, we investigate a uniquely extensive and microstratigraphically spatially uniform interval of upper Famennian (Upper Devonian) marginal marine strata in the Rocky Mountains. This transgressive deposit (< 5 m thick) rests on a depositional sequence boundary, and is composed of a thin, discontinuous basal transgressive sandstone bed, two thin fossiliferous shale beds, and several meters of oncolite-bearing carbonate wackestone. The lower shale is a consistently thin (~ 0.1 m) bed with a fossil fauna of spincicadatan (clam shrimp or conchostracans), which are extant, bivalved, chitinous, benthic crustaceans that live in fresh to brackish water, in ephemeral ponds, estuaries, and other shoreline settings. Isotopic data for the oncolitic unit record deposition in an epicontinental seaway with restricted circulation and locally brackish conditions. Sulfur isotope data may also reflect short-term changes in redox conditions, consistent with our interpretation of temporary hypoxia during deposition of the spincicadatan bed, based on its faunal assemblages. The remarkably large areal extent (~ 1600 km north to south, and ~ 1000 km east to west) of this uniformly thin interval with marginal marine fauna is one of the most unusual paleoecological events of the latest Devonian in Laurentia. It is considered an artifact of exceptional depositional processes including passive transgression (i.e., little or no wave or tidal ravinement) along with rapid opportunistic takeover of habitats during transgression. The radiation, possibly associated with temporary hypoxia, was aided by the spincicadatan reproductive strategy of numerous offspring, high growth rate, and rapid reproduction.

INTRODUCTION

Laterally extensive, relatively isochronous lithofacies and fauna have been documented from units of various ages, and they arise from a variety of depositional and stratigraphic processes that operate on a range of scales. Broad offshore settings with very low depositional slopes, e.g., the Cambrian inner detrital belt of Laurentia (Runkel et al. 2007; Myrow et al. 2012), have relatively uniform facies belts, particularly when viewed parallel to strike (Brett et al. 2007). Thin, widespread stratigraphic intervals or marker beds of relatively uniform facies and/or fauna are commonly condensed intervals formed during highstands, primarily in offshore settings (Heckel 1973, 1995; Holland et al. 2001; McLaughlin and Brett 2007).

Nearshore and shoreline settings generally produce considerable facies heterogeneity and deposits with variable and laterally discontinuous stratal geometries. This is particularly true for coastal plain settings and shoreline embayments, such as estuaries. Discontinuous facies in shoreline regions are largely a result of irregular shoreline geometries, river point sources of sediment, and along-strike variation in a number of nearshore processes including tides, waves, and currents. Shoreline environments, particularly marginal marine/coastal settings, are in many cases poorly preserved due to effects associated with shoreline migration. In particular, relative sea-level rise and coastal retreat commonly produce tidal- and wave-generated ravinement surfaces or transgressive surfaces of

erosion (Stamp 1922; Bruun 1962; Swift 1968; Helland-Hanson and Martinsen 1996; Wallace et al. 2010), which are well recognized in the rock record (Liu and Gastaldo 1992; Walker 1995; Siggerud et al. 2000; Cattaneo and Steel 2003; Laurin and Sageman 2007; Yang 2007). These surfaces record erosion and beveling of barrier island systems and marginal marine deposits, and are marked by thin overlying layers of sand (Swift et al. 1972; Hollister 1973; Pilkey et al. 1981).

Fauna in shallow settings tend to track with environments, and thus biofacies in the rock record also tend to be spatially restricted, as a reflection of the discontinuous nature of depositional systems. This is particularly true for spincicadatan or clam shrimp, which are benthic bivalved crustaceans with chitinous shells. These organisms, commonly referred to with the paraphyletic term conchostracans, live largely in freshwater, often ephemeral, ponds, but can also inhabit brackish-water settings in environments such as estuaries (Tasch 1969; Webb 1979; Vannier et al. 2003; Kozur and Weems 2010). An extraordinarily extensive, thin (<~ 0.3 m) shale layer with spincicadatan fossils exists in latest Famennian (Upper Devonian) strata within a well-defined microstratigraphy that stretches nearly the full length of the Rocky Mountains from Alberta to Nevada. We present sedimentological, stratigraphic, and geochemical data for this thin stratigraphic interval from outcrops in Utah and Montana that elucidate the origin and depositional conditions that led to its extraordinarily widespread

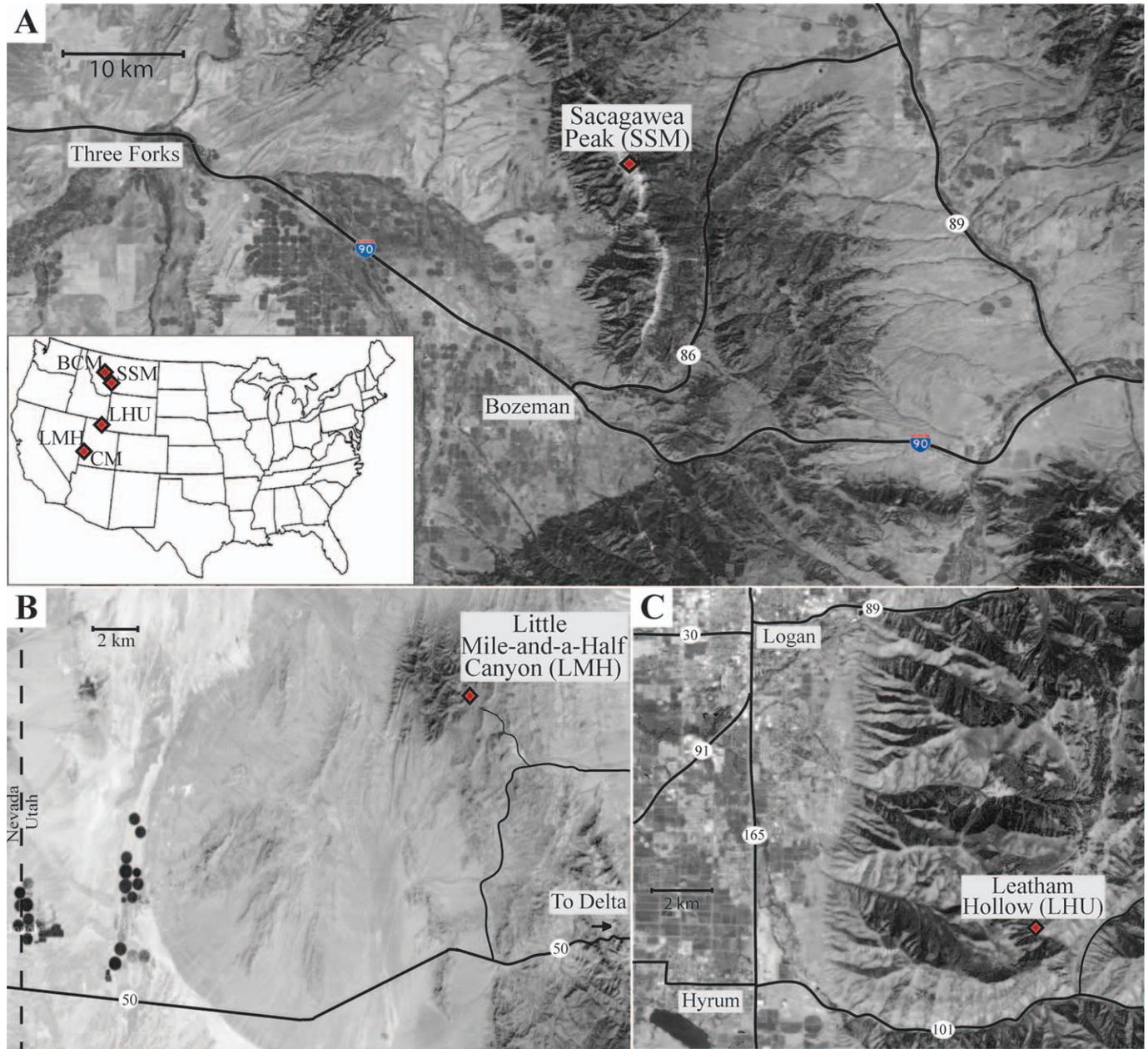


FIG. 1.—Inset map shows location of the five sites described in the paper. Close-up maps for the three main sites: A) Sacagawea Peak, Montana (SSM). B) Little Mile-and-a-Half Canyon, Utah (LMH). C) Leatham Hollow, Utah (LHU).

preservation, a reflection in part of depositional dynamics of epicontinental settings that may be applicable to other ancient deposits.

LOCATION

Data presented herein were collected from measured sections in uppermost Devonian (Famennian) strata exposed in Utah and Montana (Fig. 1). Upper Famennian strata of the Beirdneau and Leatham Formations in Leatham Hollow (LHU) are exposed southeast of Logan, Utah (N041°38.871, W111°42.824, elevation 1966 m). Approximately coeval strata of the Pilot Shale are exposed in outcrop at Little Mile-and-a-Half Canyon (LMH) in the Confusion Range, Millard Co, Utah

(N039°13.041, W113°40.124, elevation 1972 m), and at a nearby section at Conger Mountain (CM) (N039°10.948, W113°41.357, elevation 2073 m). A section of upper Famennian strata of the Three Forks and Sappington Formations was measured in the Bridger Range of Montana, ~ 25 km north-northeast of Bozeman. This section (SSM), measured along the flank of Peak 9559, next to Sacagawea Peak, is accessed through a steep valley just south of the Fairy Lake campground (N45°54.275, W110°58.593, elevation 2893 m). A section of the uppermost Sappington Formation was also measured along Beaver Creek (BCM), 35 km northeast of Helena, Montana on a steep cliff north of the road (N46°50.558, W111°45.282, elevation 1529 m).

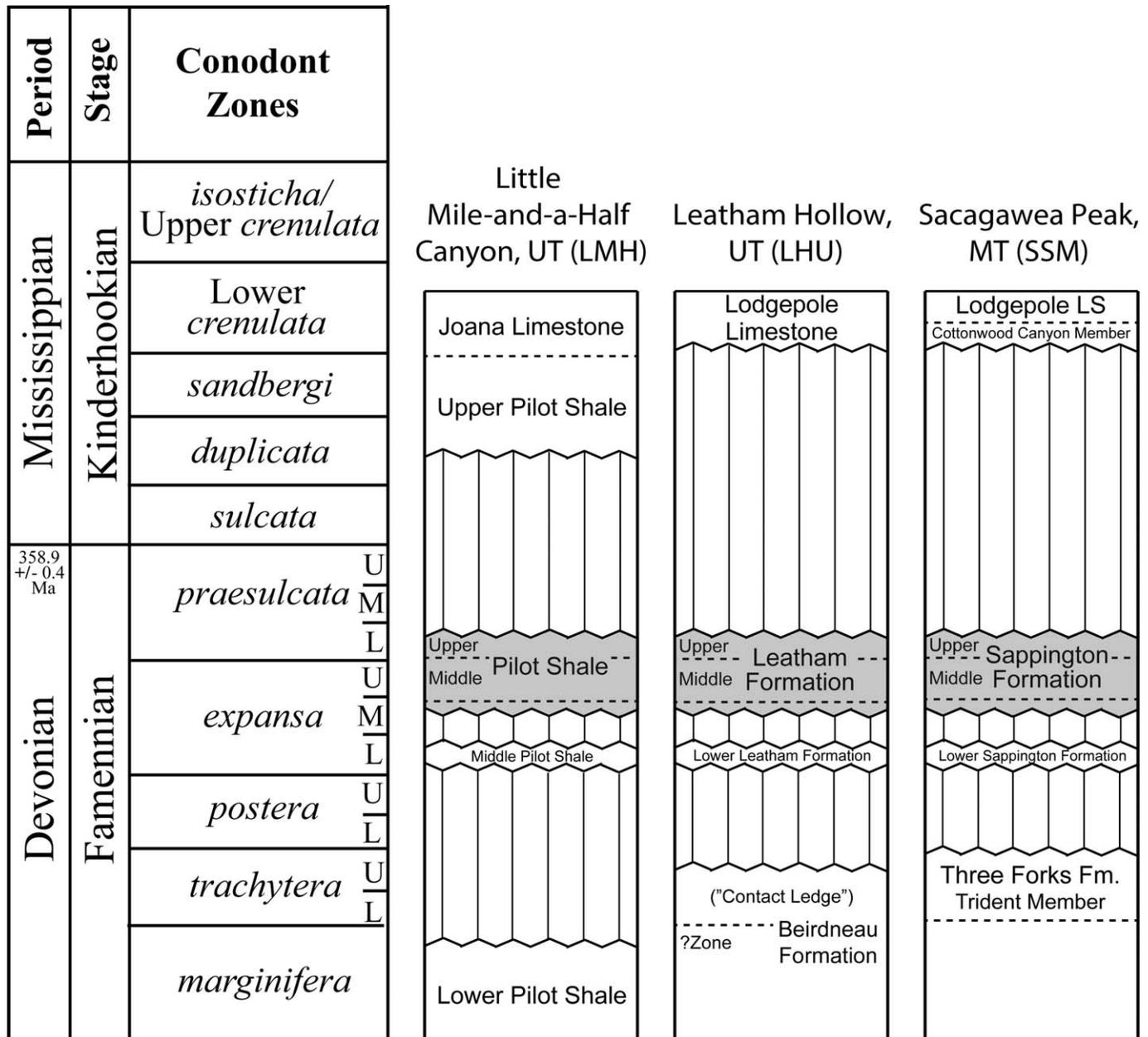


FIG. 2.—Standard Upper Devonian Conodont Zones after Sandberg et al. (2002). Date for base of Mississippian from Gradstein et al. (2012).

GEOLOGIC SETTING

The upper Famennian strata of this study represent deposition in a backbulge basin of the Antler orogeny of western Laurentia, which was in a near-equatorial position at the time (Scotese 2001). The ~ 2300-km-long Antler orogeny, which extended from the southwest United States into Canada, was caused by eastward emplacement of the Roberts Mountain allochthon, beginning in the Late Devonian (Nilsen and Stewart 1980). Flexural warping of the North American craton, as caused by thrust emplacement of the allochthon (Speed and Sleep 1982), produced the foreland basin, an upwarped forebulge, and a second wide, relatively shallow, downwarped backbulge basin. The outcrops we describe here were deposited in the backbulge basin. Changes in

accommodation space were common and spatially variable as a result of the combined effects of relative sea-level changes and shifting loci of upwarping and downwarping crust due to continued deformation. The uplifted forebulge likely created a barrier for the circulation of the epicontinental seaway filling the backbulge basin (Goebel 1991; Giles and Dickinson 1995), which likely had a greater effect during periods of low relative sea level. The barrier may have also caused stratification of the basin, as suspected from sedimentological characteristics of certain restricted lithofacies (Goebel 1991; Giles and Dickinson 1995). Transgressions and regressions recorded in these syntectonic strata resulted from a combination of both eustatic and regional tectonic effects (e.g., migration of the forebulge), processes that are difficult to separate in the rock record (Saltzman et al. 2000).

METHODS

Sedimentological, stratigraphic, and geochemical data were collected in western Laurentia where previous work had established the basic lithostratigraphic framework and conodont biostratigraphic zonation (Fig. 2). We measured and collected carbon isotope samples every ~ 0.25 m through the Pilot Shale at LMH, and through six meters of strata in the middle Pilot Shale at the CM section. At CM, samples for sulfur analysis were taken at ~ 0.5 m intervals (~ 0.1 × 0.1 m in size) through this section, along with carbonate carbon samples every ~ 0.1 m. At LHU, samples were taken every ~ 0.1 m throughout the section for carbonate carbon isotopes, and every ~ 0.25 m within the Leatham Formation for sulfur isotopes. In Montana, at the SSM and BCM sections, we collected carbon isotope samples every ~ 0.25 m.

Carbonate carbon samples were powdered using a dental drill. Approximately 150 µg samples of drilled carbonate powder were reacted with excess H₃PO₄ in He-flushed sealed tubes. Released CO₂ was then sampled using a Finnigan Gas Bench II, and isotope ratios were measured with a Delta V Plus Mass Spectrometer. Measurements are reported relative to the Vienna Pee Dee Belemnite (VPDB) standard using in-house standards calibrated with LSVEC, NBS-18 and NBS-19. Typical analytical errors were <0.1‰ (1σ) for δ¹³C_{carb} and <0.2‰ (1σ) for δ¹⁸O_{carb}. The carbon isotopic data are provided in Supplementary Data File 1.

Carbonate associated sulfate (CAS) was extracted from whole rock carbonate according to the procedure of Burdett et al. (1989), with some modifications (Gill et al. 2007). Weathered surfaces were removed from each sample using a water-cooled diamond-edge rock saw, and ~ 50 g of each sample was crushed using a SPEX 8500 shatterbox with a hardened steel grinding container. Powdered samples were rinsed in a 10% NaCl solution for a minimum of 12 hours with occasional agitation, and then rinsed twice in deionized water, each for a minimum of 24 hours. Samples were then reacted in an excess of 4N hydrochloric acid until all carbonate was dissolved, and then vacuum filtered through 0.45 micron cellulose nitrate filters. The dissolution residue was then dried and run through a chromium reduction extraction to isolate pyrite (CRS, see below). The carbonate-derived sulfate was precipitated out of the filtrate as barium sulfate, using a saturated barium chloride solution. The barium sulfate was then isolated by vacuum filtration through 0.45-micron cellulose nitrate filters. After drying, the barium sulfate was removed from the filters and boated into tin capsules with excess vanadium oxide (V₂O₅). The samples were run on a Thermo Finnigan Delta V in continuous flow mode and mated to an elemental analyzer to generate δ³⁴S.

Chromium-reducible sulfide (CRS, primarily in the form of pyrite) was extracted from CAS residue using ~ 2.0–3.5 g for each sample. Samples were loaded into a reaction vessel isolated from the atmosphere and flushed with nitrogen to carry H₂S generated from the CRS reaction into a trapping vessel. Twenty milliliters of oxygen-free, reduced chromium chloride solution was added to each digestion vessel. The digestion vessels were heated to near boiling and allowed to react for two hours as recommended by Canfield et al. (1986). The trapping vessels were filled with ~ 10 ml of 0.3M zinc acetate solution. Once the reaction was complete, the test tubes were removed from the line and 250–750 µl of silver nitrate solution was used to precipitate sulfide as silver sulfide. The samples were then centrifuged, rinsed twice with deionized water and once with 1M ammonia, then dried. The silver sulfides were then weighed into tin capsules with excess vanadium oxide (V₂O₅) and analyzed via the same mass spectrometer configuration described above. All sulfur isotopic data are provided in Supplementary Data File 2.

UPPER FAMENNIAN STRATIGRAPHY AND FACIES

A remarkably uniform and extensive, less than 5-m-thick stratigraphic succession with a thin spinicaudatan-bearing shale bed at its base exists in

upper Famennian strata throughout much of the Rocky Mountains. Strata of this interval include the middle Leatham Formation exposed at Leatham Hollow, Utah (LHU) and the upper part of the middle Pilot Shale at Little Mile-and-a-Half Canyon, Utah (LMH) (Sandberg and Pool 1977; Sandberg and Gutschick 1978; Gutschick and Rodriguez 1979). A section in the Pilot Shale at Conger Mountain (CM), near LMH, was also measured through this interval. At Sacagawea Peak (SSM) in the Bridger Range of Montana, this succession is part of the middle Sappington Formation (Gutschick et al. 1962; Gutschick 1964; Sandberg and Klapper 1967). Conodont fauna recovered from these strata are assigned to the Middle *expansa* through Lower *praesulcata* Conodont Zones (Gutschick and Rodriguez 1979; Sandberg et al. 1988; Fig. 2).

Underlying Strata

At LMH (Fig. 3), the base of the middle Pilot Shale is marked by a very thin basal sandstone and an overlying black shale-rich deposits (Sandberg and Gutschick 1969). We were unable to locate the very thin sandstone bed at LMH. The section begins with 1 m of gray calcareous shaly siltstone and then 1.7 m of slightly silty, black fissile shale (not shown in Figs. 3, 4). An overlying 0.6-m-thick covered interval (base of section in Fig. 4) is followed by a 1.7-m-thick red-brown-weathering, chert-rich interval with tabular, interbedded siltstone, very fine sandstone, black-weathering chert, and fissile black shale with beds up to 0.1 m thick. This is overlain by 1 m of chert-free, red-brown-weathering interbedded siltstone, shaly siltstone, and very fine sandstone, which mark the top of the lower part of the middle Pilot Shale (lower siltstone unit; Fig. 3).

In stratigraphically equivalent deposits at LHU, the lower Leatham Formation rests directly on the Beirdneau Formation at 15.3 m (below section shown in Fig. 4), and a 0.03–0.07-m-thick medium-coarse-grained, poorly sorted, calcareous sandstone bed marks its base. Directly above, there is 0.13 m of light-gray, brown-weathering, calcareous siltstone, followed by 0.65 m of light-gray, thinly bedded fine grainstone. The grainstone is succeeded by an ~ 8-m-thick, mostly covered interval. There are 1.10 m of poorly exposed, very thinly bedded (0.01–0.02 m), calcareous siltstone and interbedded silty shale with black chert bands within this interval at 21.6 m (Fig. 4), which we correlate to the cherty unit at LMH.

At SSM, the correlative strata of the Sappington Formation are known as the lower black shale (Fig. 4). This unit, which consists of 2.2 m of black shale, sits unconformably on an erosional surface of the underlying Trident Member of the Three Forks Formation that is highly irregular on a regional scale (Gutschick et al. 1962; Gutschick 1964; Sandberg and Klapper 1967).

Spinicaudatan-Rich Bed and Associated Strata

This fine-scale, spatially widespread stratigraphic interval is recorded in the middle Leatham Formation, upper part of the middle Pilot Shale, and in Montana the middle part of the Sappington Formation (Gutschick and Rodriguez 1979; Fig. 4). A thin (0- to 0.2-m-thick) discontinuous, poorly sorted, sandstone bed with locally abundant fish bones, *Tasmanites*, *Lumbricaria* traces, conodonts, and small phosphate pellets (Gutschick and Rodriguez 1979) marks the base of this interval (Figs. 3C, 5), where it rests directly on a black-shale-dominated interval. It was covered at LMH, and represented by an extremely thin 0.01–0.03-m-thick grainstone bed with ~ 10% fine quartz sand at section CM. A gray, brown, and black, fissile, spinicaudatan-rich shale bed up to 0.2 m thick (avg. ~0.1 m) gradationally overlies this sandstone. Conodonts recovered from this bed are assigned to the Middle *expansa* Zone (Sandberg et al. 1988). This bed is an extensive layer that is found in outcrops from Alberta to Nevada (Gutschick and Rodriguez 1979; Sandberg et al. 1980, 1988). In addition to the spinicaudatans, the bed also has fish fragments, plant spores, and marine fossils including trilobites, goniatites, Orbiculoidea, Lingula, and

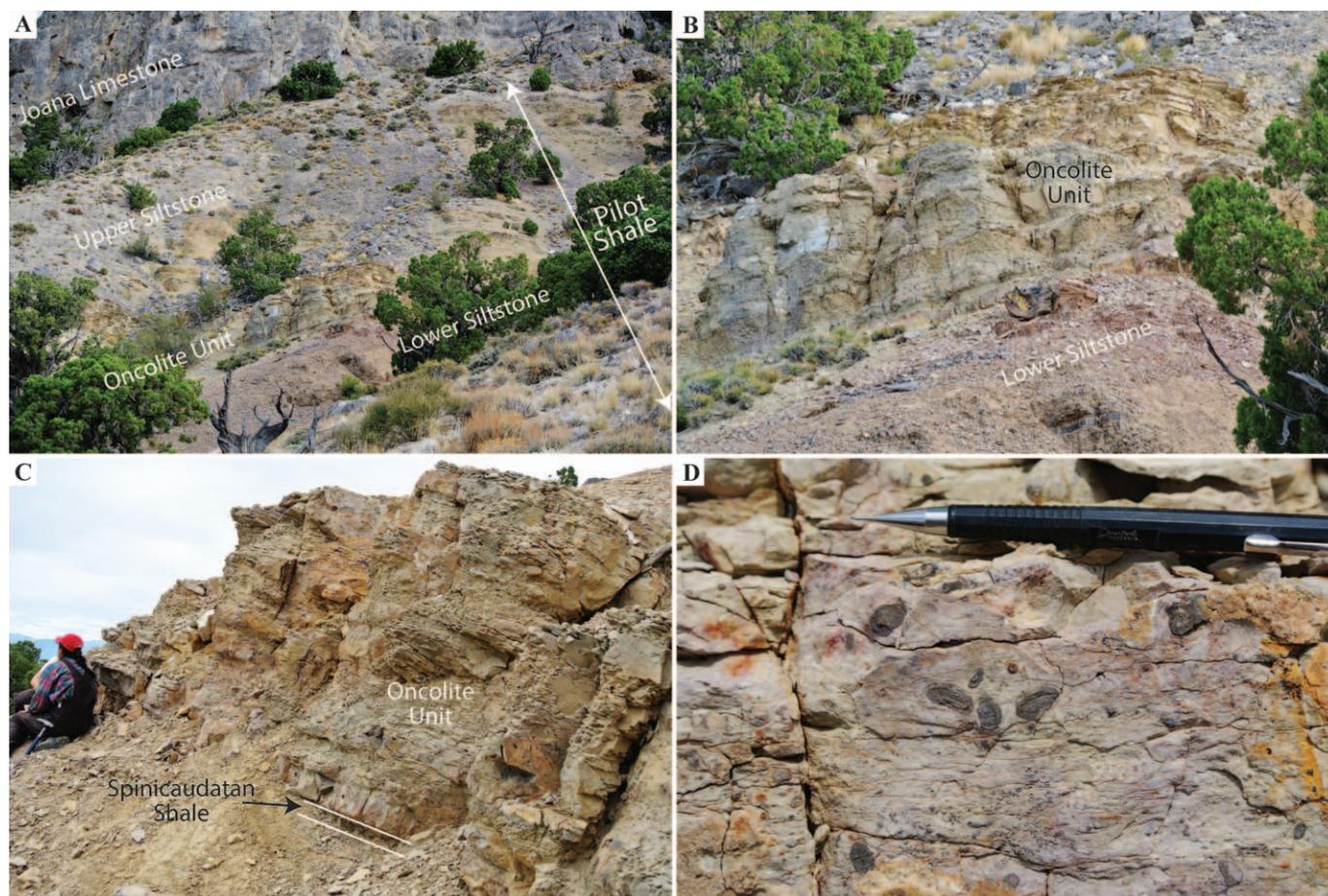


FIG. 3.—Upper Devonian to Lower Mississippian Pilot Shale exposed at Little Mile-and-a-Half Canyon, Utah (LMH). **A**) Various units of the Pilot Shale and overlying Lower Mississippian Joana Formation. Upper siltstone is 61 m thick. **B**) Oncolite-bearing lime wackestone unit (~ 3 m thick) of the middle Pilot Shale and underlying lower siltstone unit in the foreground. **C**) Closer view of oncolite-bearing unit and position of largely covered spinicaudatan-bearing shale bed. **D**) Close-up of lime wackestone with irregularly shaped oncolites. Pencil for scale.

external molds of complete skeletons of ophiuroids (Gutschick et al. 1976; Fig. 5). At SSM in Montana (Fig. 3A), this bed is overlain by 0.1–0.15 m of green-gray shale with a marine fossil assemblage of bivalves, gastropods, brachiopods, and pelmatozoan debris preserved as internal and external molds (Gutschick and Rodriguez 1979; Sandberg et al. 1980; Fig. 5). This shale unit, absent in other parts of western Laurentia, is gradational with variable carbonate-rich deposits above that include grainstone, argillaceous mudstone, wackestone, and minor calcareous shale. The remainder of the studied stratigraphic interval consists of a < 5 m thick highly variable carbonate-rich deposit that Gutschick and Rodriguez (1979) refer to as the Oncolite-Shell-Sponge Bank Unit (Figs. 3, 6B, C). Abundant oncolites have fossil nuclei, usually brachiopods (e.g., *Rhipidomella missouriensis*), and concentric lamination that enveloped foraminifera, coral, and bryozoan epifauna (Gutschick et al. 1962). This unit is generally assigned to the Upper *expansa* Conodont Zone. However, in two locations in Utah, the uppermost beds of this interval have yielded a Lower *praesulcata* conodont fauna (Sandberg et al. 1988).

The base of the oncolite-bearing unit at LMH consists of 0.62 m of grainstone with locally abundant bioclasts and both oncolites and concretionary nodules that are highly irregularly shaped (nonspherical). The grainstone is interbedded with shaley carbonate mudstone with irregular carbonate nodules. This interval is overlain by 2.8 m of 0.2–0.5-m-thick beds of carbonate wackestone with dispersed irregular

nodules, interbedded with 0.05–0.3 m beds of fine grainstone. This oncolite unit contains an abundant and diverse fossil assemblage of foraminifera, brachiopods, bryozoans, calcareous sponges, goniatite cephalopods, bivalves, gastropods, ostracods, holothuroidean sclerites, trilobites, conodonts, fish fragments, and a variety of trace fossils (Gutschick and Rodriguez 1979; Fig. 5). A 2.9-m-thick correlative section was measured in the middle Pilot Shale near LMH at Conger Mountain (CM). This section has a 1–3-mm-thick, limonite stained, fine grainstone with 1%–10% fine to coarse quartz sand at the base of the oncolite-bearing unit. Carbonate mudstone, wackestone, fine peloidal calcisiltite, massive, gray, fine grainstone, and limonite-stained marl beds with abundant fossils and large oncolites (up to 4 cm diameter) overlie this extremely thin bed.

The stratigraphic succession at LHU has a 0.1-m-thick bed of black calcareous shale at its base, and an overlying 0.12 m bed of black carbonate mudstone with sparse spinicaudatan fossils. We correlate this basal black shale bed to the thin spinicaudatan-rich black shale unit described above, although we did not locate the thin underlying sandstone bed as described by Sandberg and Gutschick (1969). The “oncolite-bearing” carbonate unit at this locality lacks the large oncolites present elsewhere in the Rocky Mountain region. It has a basal 0.25-m-thick gray, coarse, crinoidal grainstone bed, whereas the rest of this unit (3.41 m thick) consists of slightly fissile, light-gray, thick-bedded (0.42–0.80 m), slightly shaley carbonate mudstone with irregular to flattened carbonate nodules.

Little Mile-and-a-Half Canyon, UT (LMH)

Leatham Hollow, UT (LHU)

Sacagawea Peak, MT (SSM)

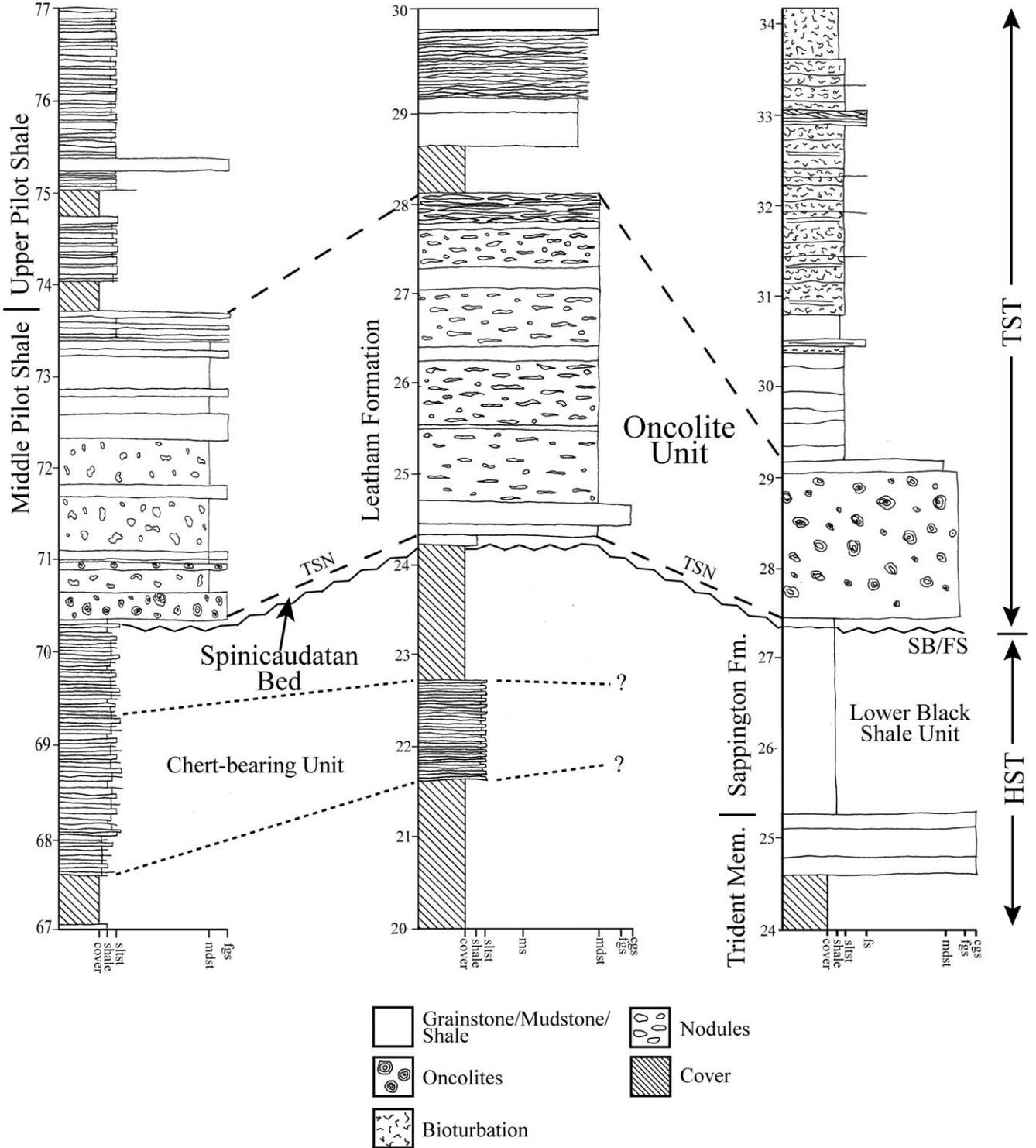


FIG. 4.—Detailed stratigraphic sections for the studied interval at Little Mile-and-a-Half Canyon, Utah (LMH); Leatham Hollow, Utah (LHU); and Sacagawea Peak, Montana (SSM). Abbreviations: Siltst = siltstone; mdst = carbonate mudstone/wackestone; fs = fine sandstone; fgs = fine grainstone; cgs = coarse grainstone; SB/FS = Sequence boundary/Flooding surface; TSN = Transgressive surface of nonerosion, which records a shift from marginal marine to fully marine conditions; HST = Highstand systems tract; TST = Transgressive systems tract.

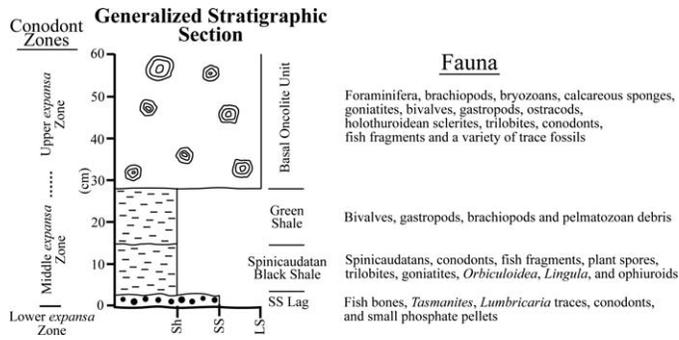


FIG. 5.—Generalized stratigraphic section of the basal part of the studied interval and its component fauna. Faunal lists from Gutschick and Rodriguez (1979) and conodont biostratigraphy from Sandberg et al. (1988).

At SSM, the base of the studied succession is marked by 0.01 to 0.02 m of indurated black shale with brachiopods and spinicaudatan fossils. The shale is overlain by 1.6 m of an oncolite-bearing unit of petroliferous, yellow-weathering, mottled-gray, calcisiltite to fine grainstone with abundant crinoid and brachiopod bioclasts (~ 15%–20%) (Fig. 6B, C). The upper 80 cm is more argillaceous and flaggy weathering. Individual oncolites in this unit are up to 0.05 m in diameter. This unit is overlain by 1.8 m of yellow-weathering calcareous siltstone with 5%–10% dispersed bioclasts, including brachiopod and crinoid fragments.

Above the studied succession in Utah, the upper Pilot Shale is a poorly exposed slope-forming unit of tan- to yellow-weathering, interbedded calcareous siltstone, silty shale, and minor beds of fine to very fine silty grainstone. The age-equivalent “upper siltstone” of the upper Sappington Formation in Montana consists dominantly of shale and siltstone, although it contains more sandstone with parallel and wave-ripple lamination in its upper part. At Beaver Creek, Montana, this unit contains abundant trace fossils, primarily the small highly curved *Phycosiphon incertum* and the meniscate trail *Nereites missouriensis* (identifications by L. Buatois, personal communication 2013) (Fig. 6D). A black shale unit caps the upper Pilot Shale in sections in Montana.

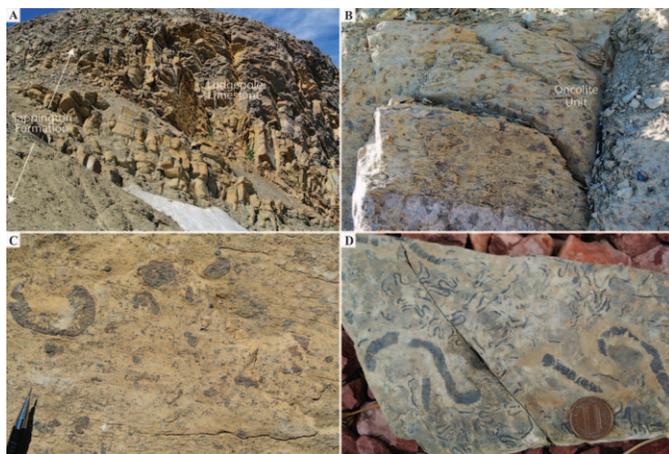


FIG. 6.—A) Upper Devonian Sappington Formation (~ 45 m exposed in foreground) and overlying Mississippian Lodgepole Limestone near Sacagawea Peak (SSM) in the Bridger Range, Montana. B) Oncolite-bearing unit at SSM. Pencil for scale in crack on right. C) Close-up of wackestone with abundant irregular oncolites and skeletal debris. D) Trace fossils *Phycosiphon incertum* (the smaller contorted traces) and the meniscate trail *Nereites missouriensis* from the upper siltstone part of the Sappington Formation.

Spinicaudatan and Other Associated Fauna

Spinicaudatans (clam shrimp), also known by the paraphyletic term conchostracans, are bivalve, weakly biomineralized, extant branchiopod crustaceans. The extinct species present in this study is *Cyzicus lioestheria*, which has been reported in the fossil record from the Upper Devonian to the Upper Cretaceous (Tasch 1969; Gutschick and Rodriguez 1979). The majority of spinicaudatans, especially extant varieties, are known to live in small, often ephemeral, freshwater ponds (e.g., playa lakes), but can also inhabit brackish-water settings in environments such as estuaries, embayments in streams, salt flats, coastal rock pools, and deltaic plains (Tasch 1969; Webb 1979; Vannier et al. 2003; Olempska 2004; Stigall and Hartmann 2008; Kozur and Weems 2010). Such species may have been more prevalent in the Late Devonian, as the ancestors of the spinicaudatans were marine, and their evolution into the occupation of freshwater niches likely occurred over the course of the Devonian (Vannier et al. 2003; A.L. Stigall personal communication 2014). A few post-Devonian examples have been found in association with marine fossils, but in extremely shallow water lithofacies and with the possibility of mixing of fauna during transport (Webb 1979).

Modern members of the cyzicid family of spinicaudata are relatively resistant to temperature change, moving freely through stratified waters, although they appear to function most efficiently at temperatures above 8°C (Eriksen and Brown 1980). Additionally, cyzicids have the ability to survive relatively low oxygen concentrations, with a low rate of oxygen consumption relative to their body weight in comparison to other phyllopods (Eriksen and Brown 1980). In the majority of species, eggs are not only able to withstand remarkable environmental stress including desiccation, freezing, and anoxia, but require drying and rehydration to hatch (Webb 1979; Frank 1988; Thiery 1996; Vannier et al. 2003). Thus, sudden flooding events can initiate synchronous hatching of a vast population, which can be recorded in the fossil record as a bedding plane assemblage of organisms of similar size and maturity (Stigall and Hartman 2008).

FACIES INTERPRETATIONS

Cole et al. (2015) interpret the strata of the lower Leatham Formation (LHU), lower to middle Pilot Shale (LMH), and lower Sappington Formation (SSM), which rest below our studied interval, to represent a full depositional sequence. These underlying Lower *expansa* Zone strata have a thin sandstone bed at the base and top, both of which mark sequence boundaries; the lower sequence boundary has a large diastem (missing *postera* Zone at LHU, and missing *postera* and *trachytera* Zones at LMH). The extremely thin and discontinuous sandstone bed at the top of these older strata, which defines the base of our studied units (middle Leatham Formation, upper part of middle Pilot Shale, and middle Sappington Formation), separates Lower *expansa* from Middle *expansa* Zone strata. The stratigraphically equivalent sandstone bed and sequence boundary is not well exposed at SSM, but it is present in nearby areas within Montana (Sandberg and Poole 1977; Sandberg and Gutschick 1978; Gutschick and Rodriguez 1979). The sandstone bed at the base of our studied interval rests directly on a sequence boundary and the highstand systems tract of the underlying sequence (Fig. 4). The sandstone may represent a very thin transgressive lag, based on its stratigraphic position and presence of reworked fish bones.

The overlying black shale bed with the spinicaudatan *Cyzicus lioestheria* was deposited in extremely shallow water marginal marine conditions, given the ecological distribution of modern and ancient spinicaudatans described above as well as the mixed assemblage of fish fragments, plant spores, and varied marine invertebrates. The presence of both articulated spinicaudatans and the molds of articulated starfish skeletons argue for extremely quiet water conditions (Gutschick 1964; Gutschick et al. 1976). The stratigraphic transition from this shale with

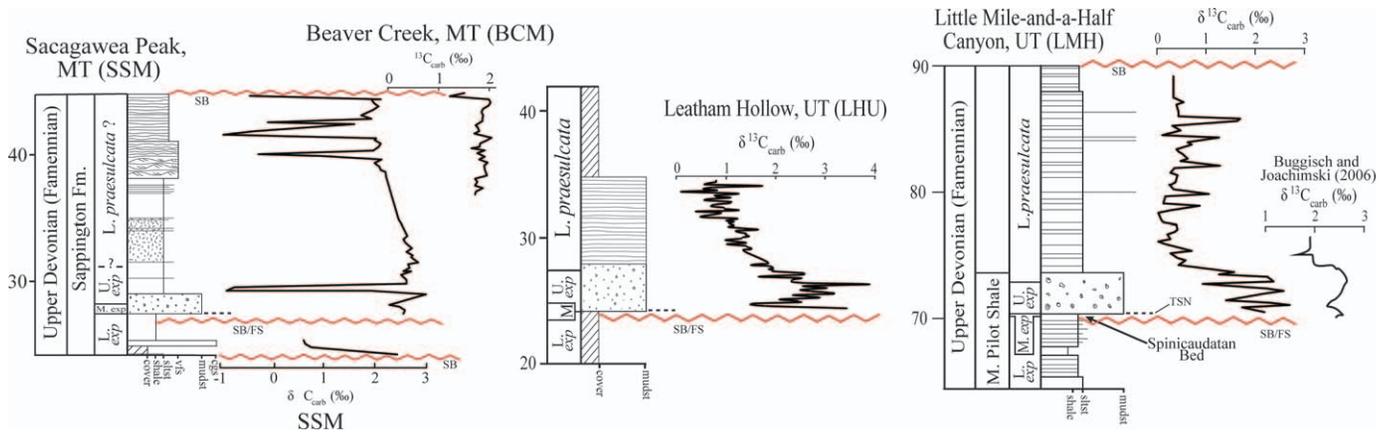


FIG. 7.— $\delta^{13}\text{C}_{\text{carb}}$ —data for Upper Devonian strata of the Sappington Formation at Sacagawea Peak, Montana (SSM) and Beaver Creek, Montana (BCM), and strata of the middle upper Pilot Shale at Little Mile-and-a-Half Canyon, Utah (LMH) and Leatham Hollow, Utah (LHU) with associated biostratigraphic data. $\delta^{13}\text{C}_{\text{carb}}$ data for Upper *expansa* to basal Lower *praesulcata* Zone strata from Buggisch and Joachimski (2006) is shown on far right. For symbols see legend for Figure 4.

mixed marginal marine and fully marine fauna to overlying strata with solely marine fauna, suggests the possibility of deposition in restricted, semi-enclosed, shoreline settings, including possibly brackish environments. The prasinophyte alga *Tasmanites*, which is associated with the spinicaudatan-bearing bed of this study, has been found in marine, brackish, and freshwater environments (Telnova 2012).

As mentioned earlier, this bed has been identified within the thin, yet remarkably extensive unit stretching from Alberta to Nevada (Gutschick and Rodriguez 1979; Sandberg et al. 1980). The stratigraphic transition to the overlying shale bed with marine fauna at SSM (Gutschick and Rodriguez 1979; Sandberg et al. 1980), and evidence of the mixing of marginal marine and marine faunas in the spinicaudatan bed, suggests transgression and a possible transition from low salinity or euryhaline conditions to a more stable shallow-marine environment. The extensive nature of this thin interval suggests passive transgression, with little or no erosion over vast distances.

The overlying carbonate unit contains a wide variety of open-marine fossils, and based on numerous closely spaced sections, it was a carbonate buildup with meters of synoptic relief on the seafloor (Gutschick and Rodriguez 1979). The generally abundant, relatively large oncolites, and the coarse grainstone at the base of the unit at LMH, indicate deposition under high-energy conditions. Those parts of the carbonate unit with oncolites dispersed within carbonate wackestone suggest transport of the oncolites from an adjacent higher-energy environment.

Several lines of evidence suggest at least temporary hypoxia during deposition of the lowermost part of the study interval. First, the dark, organic rich nature of the shale, along with a pyritized fauna at the transition into the basal part of the oncolite unit, supports low oxygen conditions (Myrow 1990). Second, the fauna in the bed are diminutive (Gutschick and Rodriguez 1979), which is consistent with modern settings with seasonal hypoxia in which the fauna are small to better utilize resources (Levin et al. 2009). Third, the presence of abundant *Tasmanites* prasinophyte algae in the sandstone interval at the base of the spinicaudatan bed is also consistent with hypoxia because prasinophytes bloom by effectively utilizing reduced nitrogen species (Eppley et al. 1969; Prauss 2007). Prasinophytes exist in the rock record in great abundance in conjunction with organic-rich shale, and are associated with mass mortality and extinction of other marine organisms, i.e., they are commonly a disaster taxon (Tappan 1980). They are often found at the base of transgressive systems tracts in association with transgressive lags, as in this study, presumably because nutrients such as reduced nitrogen or phosphate residing in marine pore waters are remobilized during

transgression, or are brought into the environment by upwelling bottom water (Prauss 2007). In either case, the increase in primary productivity leads to consumption of oxygen and eutrophication. Gutschick and Rodriguez (1979) describe phosphate nodules in the sandstone lag below the spinicaudatan bed, supporting the idea that high nutrients led to algal blooms and eutrophication. Concentrations of fish bones in the sandstone lag are also consistent with low oxygen conditions, or even euxinia (Caswell and Coe 2012).

The spinicaudatans may represent a largely opportunistic fauna (e.g., Diaz 2001) brought on by temporary hypoxia, but with other marine fauna mixed in by immigration during periods of normal marine water, and in cases possibly by physical transport. This situation would potentially be analogous to population explosions of highly abundant opportunistic epifaunal bivalves spurred by increases in primary production, specifically *Tasmanites* blooms, following Early Jurassic oceanic anoxic events (Caswell and Coe 2013). Modern spinicaudatans have many of the characteristics of opportunistic taxa (Levin et al. 2009) including an r-type selection reproductive strategy with type III survivorship patterns, meaning large larval broods and low percentages of survival, as well as short lifespans, high growth rates, and small body sizes (Tasch 1969; Webb 1979). In addition, the spinicaudatan-rich shale bed may also record, in part, a series of mass mortality events possibly brought on by temporary anoxia.

The exact position(s) of a marine flooding surface(s) is difficult to pinpoint, although the extremely thin sandstone lag and associated thin spinicaudatan-rich bed in general represents marginal marine conditions and initial marine onlap. The transitions to marine shale and then the overlying oncolite unit represent one or more additional marine flooding surfaces that had little or no erosion associated with them. Below we explore the geochemical signature of this remarkably widespread passive transgressive succession (lowermost transgressive systems tract), in part to determine the nature of transitions in bottom-water chemistry in these strata, as they might relate to the deposition of these widespread marginal marine and marine Upper Devonian lithofacies and fauna.

CARBON ISOTOPE GEOCHEMISTRY

$\delta^{13}\text{C}_{\text{carb}}$ Results

The Lower to Middle *expansa* Zone spinicaudatan-bearing black shale interval in Utah lacked carbonate and thus was not suitable for carbon isotope analysis. The lowest samples of the studied interval at LMH and LHU were taken from the base of the oncolite-bearing carbonate unit in

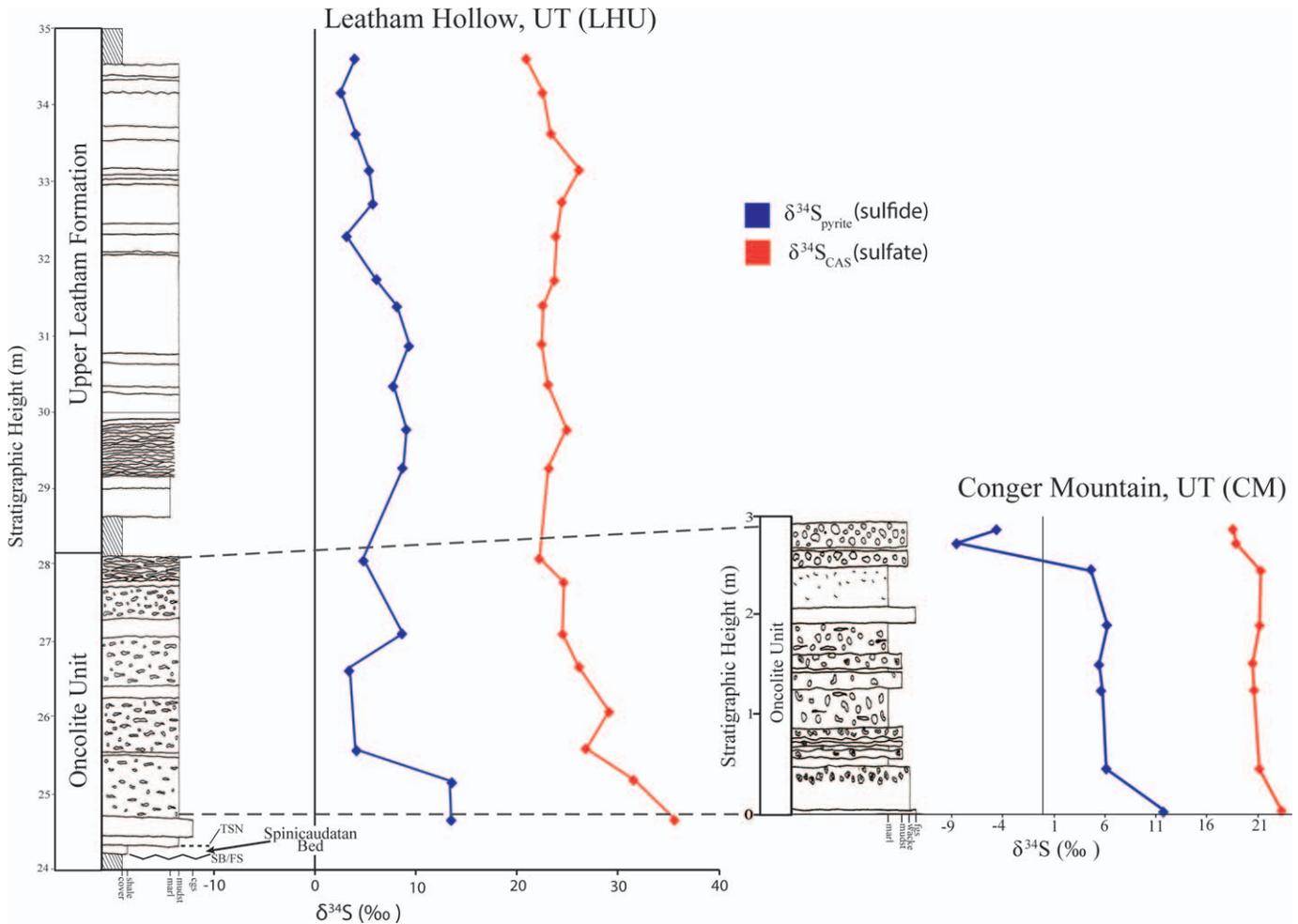


FIG. 8.— $\delta^{34}\text{C}_{\text{pyrite}}$ (sulfide) and $\delta^{34}\text{C}_{\text{CAS}}$ (sulfate) isotopic data for Leatham Hollow, Utah (LHU) and Conger Mountain, Utah (CM). For symbols see legend for Figure 4.

Upper *expansa* strata (Fig. 3). $\delta^{13}\text{C}_{\text{carb}}$ data are presented for Upper *expansa* through Lower *praesulcata* strata from the middle and lowermost-upper Pilot Shale at LMH, the middle and upper Leatham Formation at LHU, and the middle and upper Sappington Formation at SSM (Fig. 7).

At LMH, $\delta^{13}\text{C}_{\text{carb}}$ values initially oscillate between 1.0‰ and 2.8‰ with the average decreasing from $\sim 2.2\%$ to $\sim 1.3\%$, then abruptly shift to lower values at the base of the Lower *praesulcata* Zone. Above, values are consistent and average $\sim 0.5\%$, except for a few more positive values between 0.9‰ and 1.7‰, which may have resulted from the presence of numerous nodules that may not record true secular variations in the marine isotope record. $\delta^{13}\text{C}_{\text{carb}}$ values at LHU show similarly positive values and variation (1.5‰ to 3.9‰) through the oncolite bank, and a similar shift to decreasing values in the upper part of the Leatham Formation at the base of the Lower *praesulcata* Zone. $\delta^{13}\text{C}_{\text{carb}}$ values in the oncolitic carbonate unit and the overlying 10 m of strata at SSM are $\sim 2\%$ – 3% , with the exception of two anomalous negative values close to -1% just above the oncolite unit. The uppermost Lower *praesulcata* Zone strata at SSM have similar values that average $\sim 2\%$ but with scattered negative values that reach -1.0% . At BCM, values for the same interval also average $\sim 2\%$, but show little variability.

Interpretations: $\delta^{13}\text{C}_{\text{carb}}$

A few $\delta^{13}\text{C}_{\text{carb}}$ values from Lower *expansa* Zone strata at SSM, which underlie our studied stratigraphic interval, include one from a thin carbonate bed at 24.24 m (2.5‰), and three overlying beds with sample values that average $\sim 0.7\%$ (Fig. 7). Such values are more positive than those of underlying *marginifera* and *trachytera* Zone strata, and reflect the beginning of a long-term positive shift associated with late Famennian to Mississippian glaciation (Cole et al. 2015). The Lower *expansa* Zone data are similar to those for the same age strata elsewhere (Myrow et al. 2011). There are too few data to determine if an isotopic signal of the Lower *expansa* Zone Dasberg event is present (Kaiser 2005; Buggisch and Joachimski 2006; Kaiser et al. 2008; Myrow et al. 2011). A shift to Middle *expansa* Zone at the basal sandstone of the studied interval is consistent with a global regression at the end of the Dasberg event (Becker 1993; Sandberg et al. 2002; Kaiser et al. 2006, 2008; Hartenfels and Becker 2009).

$\delta^{13}\text{C}_{\text{carb}}$ isotope values generally close to 2.0‰ are typical of Upper *expansa* Zone strata (Buggisch and Joachimski 2006), and the $\sim 1\%$ – 2% negative shift at the upper *expansa*–*praesulcata* boundary at LMH and LHU (Fig. 7) correlate well to a similar shift in European strata at this level (Buggisch and Joachimski 2006; Kaiser et al. 2006). The 1.5‰ and 2.0‰

$\delta^{13}\text{C}_{\text{carb}}$ values are consistent with those of Lower and Middle *praesulcata* Zone strata (Buggisch and Joachimski 2006; Kaiser et al. 2006).

More importantly, the samples from the lowermost part of the oncolite unit do not show abnormal values (i.e., highly negative), which would reflect deposition or early diagenesis in brackish or nonmarine fluids. Thus, the transition from marginal-marine to marine conditions (i.e., marine flooding surface) likely took place at the transition at the top of the spinicaudatan-bearing shale bed.

Sulfur Isotope Geochemistry: Results

Leatham Hollow.—The sulfur isotope values from pyrite (CRS) range from 4‰–14‰, with most variability in the lower part of the sampled section (Fig. 8). Sulfide $\delta^{34}\text{S}$ is highest (13.5‰) in the basal oncolite bank. At 25.56 m, values drop to 4.1‰, then return to 8.8‰ through the overlying ~2 m. Between 29.24 and 31.69 m, values are fairly consistent with minimal oscillation around ~8.0‰. There is a trend toward slightly lower values (4.3‰) toward the topmost part of the sampled section.

There is also significant variation in the isotopic composition of carbonate-associated sulfate (CAS) ($\delta^{34}\text{S}_{\text{CAS}}$ from 20.8‰–35.5‰), although there is the same generally decreasing upward trend through the sampled section (Fig. 8). At the base of the section, $\delta^{34}\text{S}_{\text{CAS}}$ values are 35.5‰ and drop to 26.1‰ within the lowermost 2 m. Above this point, values remain between ~22.0‰ and ~26.0‰, up to the topmost sample, which has a value of 20.8‰.

The average difference between the sulfur isotope values for sulfate and sulfide is 18.0‰. However, this fractionation is consistently less through the middle of the section (27.07–31.35 m), with the lowest value (13.0‰) at 30.85 m.

Conger Mountain.—The sulfur isotope values from sulfide at Conger Mountain cover a larger range (–8‰ to +12‰) than in Leatham Hollow, but the majority of this variability is at the top and bottom of the sampled section (Fig. 8). Sulfide $\delta^{34}\text{S}$ in the middle part of the section has little variation. Sulfide $\delta^{34}\text{S}$ values at the base begin at 11.9‰, but drop within 0.50 m to 6.2‰. There is little variation through the next 2 m, with values ranging between 6.2‰ and 4.7‰. The samples from the top 0.50 m of the section show a negative shift to values of –8.5‰ and –4.6‰.

The $\delta^{34}\text{S}$ values from CAS are consistent throughout this section, with a slight negative shift at the base and top of the section. These shifts coincide with the changes in sulfide $\delta^{34}\text{S}$ and are of smaller magnitude (Fig. 8). Values through the middle of the section are between 20.5‰ and 21.3‰, and drop to ~18.7‰ in the top 0.50 m.

Sulfur isotope offsets between sulfide and sulfate show some co-variation, with an average of 17.6‰ in the section. The difference between the sulfur isotope values for sulfate and sulfide averages 17.6‰ throughout the section. This fractionation generally increases upsection from a minimum of 11.5‰ in the lowermost sample, to an average of 15.0‰ through the middle part of the section, increasing to ~25.0‰ in the topmost 0.50 m.

Interpretations: $\delta^{34}\text{S}$

The isotopic composition of carbonate-associated sulfate ($\delta^{34}\text{S}_{\text{CAS}}$) is generally thought to track that of seawater sulfate (Kampschulte and Strauss 2004; Lyons et al. 2004), and thus serves as both a record of the ancient marine sulfur cycle and as a tool for chemostratigraphic correlation (Fike and Grotzinger 2008). However, $\delta^{34}\text{S}_{\text{CAS}}$ can diverge from the seawater $\delta^{34}\text{S}_{\text{SO}_4}$ trend as a result of both primary environmental variability (e.g., carbonate deposition in a brackish environment or a restricted basin with sulfate composition not representative of open ocean seawater) or as the result of diagenetic alteration (Marenco et al. 2008). Here, we observe a general trend of decreasing $\delta^{34}\text{S}_{\text{CAS}}$ values in

both Conger Mountain and Leatham Hollow. However, there is a lateral difference in $\delta^{34}\text{S}_{\text{CAS}}$ between these sections, with $\delta^{34}\text{S}_{\text{CAS}}$ in CM ~4‰ lower than in LHU. In addition, there is little stratigraphic variability in CM, while there is substantially higher (~14‰) variability in LHU. Taken together, the higher $\delta^{34}\text{S}_{\text{CAS}}$ values in LHU and their increased stratigraphic variability suggest that they have been altered relative to $\delta^{34}\text{S}_{\text{CAS}}$ preserved in CM. The slightly heavier overall values could be due to increased microbial sulfate reduction (MSR) within the basin that locally increased the $\delta^{34}\text{S}$ composition of the water, or due to alteration of the sulfate pool within the upper part of the sediment column (e.g., during lithification).

Both CM and LHU strata were deposited in an epicontinental seaway that previously published work suggests was significantly restricted through much of this period (Gutschick and Rodriguez 1979; Sandberg 1980; Sandberg et al. 1988). However, there are few published data of sufficient resolution for comparison. Existing $\delta^{34}\text{S}_{\text{sulfate}}$ data from the Nisku Formation in Alberta (Machel 1985) show similar variability and absolute values (averaging ~20‰–25‰) as the present study; however, the Nisku Formation was also likely deposited in an epicontinental seaway, rather than representative of the open ocean. In comparison, CAS from rocks of this general age, which are interpreted to represent the open ocean, have values averaging ~23‰ (Kampschulte and Strauss 2004). These previously published results agree with the lower $\delta^{34}\text{S}_{\text{CAS}}$ values (i.e., from CM and the upper part of LHU) presented here, suggesting these may track the regional seawater composition, while the higher $\delta^{34}\text{S}_{\text{CAS}}$ values at the base of LHU likely reflect local alteration. While lateral gradients in water chemistry are plausible given the inferred brackish nature of deposition of the spinicaudatan-rich strata, and are consistent with the carbon isotopic heterogeneity found in these strata across the basin, this is difficult to uniquely diagnose with $\delta^{34}\text{S}$ (Fig. 8; Cole et al. 2015).

The $\delta^{34}\text{S}_{\text{sulfide}}$ values, which range from –9‰ to 14‰, are higher than expected for average marine pyrites at this time (Canfield 2001; Sageman et al. 2003). Fractionation during this time is estimated to be ~30‰ (Wu et al. 2010), whereas our fractionations average 17.1‰. In modern sediment, high $\delta^{34}\text{S}$ values in sulfides (i.e., small fractionations) can be produced in a variety of settings (Leavitt et al. 2013; Wing and Halevy 2014), including shallow-marine environments that are controlled by physical reworking (Aller et al. 2010). These systems undergo significant oxidative and reductive sulfur cycling, and are characterized by elevated $\delta^{34}\text{S}_{\text{pyr}}$ values (reaching up to +21‰ in modern sediment). Strongly varying redox conditions in these environments also can induce local changes in microbial sulfur cycling and associated changes in the isotopic fractionation between coeval sulfate and sulfide (Canfield and Teske 1996; Canfield 2001; Sim et al. 2011). Our interpretation of possible temporary hypoxia during deposition of the spinicaudatan bed is consistent with such varying redox conditions, and the general interpretation of possible restricted circulation during deposition of these units.

These factors (controls on sulfate reduction; importance of sulfide oxidation; and partial closed-system behavior) control the observed fractionation between sulfate and sulfide. The stratigraphic patterns in the fractionation between sulfate and sulfide suggest variable importance of these factors throughout the section. The lowest $\delta^{34}\text{S}_{\text{sulfate}} - \delta^{34}\text{S}_{\text{sulfide}}$ fractionations are most likely the result of either physical factors, such as an increase in sulfide oxidation related to syndepositional reworking, a deeper chemocline, or changes in the conditions of sulfate reduction in the sediment. These variables are difficult to untangle, and it is likely that a combination of these processes is responsible for the stratigraphic patterns and variability of our isotopic data. Finally, the lateral gradient between the $\delta^{34}\text{S}_{\text{CAS}}$ values at LHU and CM is most pronounced through the oncolite bank, indicating the most significant local alteration during the deposition of these strata. Unfortunately, the salinity of the system,

especially for the underlying spinicaudatan-rich bed cannot be deciphered from our geochemical data.

DISCUSSION

The remarkably widespread, fine-scaled stratigraphy of the studied interval, from basal thin sandstone through oncolite-bearing carbonate, reflects consistent depositional conditions over much of this part of western Laurentia at this time. The north-south extent of the thin spinicaudatan-bearing shale bed is ~1600 km, from southern Nevada to southern Alberta. The east-west extent, from southeastern Nevada and Alberta to central North Dakota, including the Bakken Formation of western North Dakota (Thrasher 1987) and the Exshaw Shale of western Alberta (Macqueen and Sandberg 1970), is ~1000 km. Such marginal marine lithofacies and fauna are not typically extensive at the orogen scale.

Basal transgressive deposits commonly represent high-energy environments and consist of well-sorted sandstone units that are extensive, but often highly irregular in thickness (e.g., Myrow et al. 2003). The thin, discontinuous sandstone bed at the base of our studied interval, which rests on a thick shale-rich interval devoid of sandstone beds, is consistent with a shift from Lower to Middle *expansa* Zone conodonts at the basal sandstone, and thus the presence of a sequence boundary and overlying transgressive lag. The chronostratigraphic position of the sequence boundary is also consistent with a global regression at this time, namely the last phase of the Dasberg event (Becker 1993; Sandberg et al. 2002; Kaiser et al. 2006, 2008; Hartenfels and Becker 2009).

In most cases, the potential for preservation of marginal marine strata in basal transgressive systems tract deposits is very low due to reworking by shallow-marine processes. In settings characterized by moderate to strong tidal and/or wave processes, transgressive surfaces of erosion (i.e., ravinement surfaces) form (e.g., Posamentier and Vail 1988; Dalrymple et al. 1992, 1994; Allen and Posamentier 1993; Galloway 2001), and these tend to remove very shallow water deposits. The stratigraphic transition above the disconformity surface at the base of the discontinuous thin sandstone bed shows no evidence of either wave- or tidal-generated sedimentary structures, e.g., cross-bedding. The extensive deposition of a thin marginal marine shale bed, overlying thin marine shale bed, and capping oncolite-bearing carbonate deposit suggest upward deepening from a quiet water marginal marine setting to a more energetic carbonate environment. Our $\delta^{13}\text{C}_{\text{carb}}$ data from the oncolitic unit show typical marine values, not highly negative values that might reflect either deposition or early diagenesis in brackish or nonmarine fluids. Thus, the position of the marine flooding surface within this transgressive systems tract deposit rests near or at the top of the spinicaudatan-bearing shale bed.

$\delta^{34}\text{S}_{\text{CAS}}$ data are not available for the marginal-marine deposits of the spinicaudatan-rich bed, so no comparison can be drawn with overlying strata in this data set. $\delta^{34}\text{S}_{\text{CAS}}$ values in the overlying oncolite unit are highly variable, and in some sections slightly heavier overall values could reflect basinal increases in microbial sulfate reduction (MSR), local increases in the $\delta^{34}\text{S}$ water composition, or early diagenetic alteration of the sulfate pool within the upper sediment column. Some of the variability may be due to deposition within a generally restricted basin where the sulfate pool was not fully buffered, thus allowing for large short-term changes in $\delta^{34}\text{S}_{\text{sulfate}}$ values. The sulfur isotope data may in part reflect short-term variations in redox conditions, consistent with the unusual faunal composition of the spinicaudatan bed and our resulting interpretation of temporary hypoxia during its deposition.

The large spatial extent of the fine-grained spinicaudatan-bearing bed of this study, and the limited environmental/paleoenvironmental distribution of modern and ancient spinicaudatans, namely extremely shallow, fresh, and brackish-water settings, is unusual for much of the rock record. In particular, marginal marine fauna and associated facies tend to have significant along-strike variability, which if migrated landward during

a transgression would tend to produce discontinuous stratigraphic units and facies mosaic patterns, even if fully preserved. Extensive thin shale intervals with uniform thickness and fauna are commonly offshore deposits with open marine facies that represent condensed intervals, in contrast to the paleoenvironments of the units in this study.

The initial transgressive units were deposited over widespread areas both parallel to and across the length of the basin margin. Although fairly precise in chronostratigraphic terms, there is no way to resolve the temporal history of colonization of the environments recorded by the spinicaudatan fossils, which had a very narrow environmental tolerance. The spatial distribution of these Devonian fauna suggests that a widespread marginal marine colonization by these crustaceans tracked rising sea level over a geologically relatively short time interval (Middle *expansa* Zone) during a relatively passive transgression. The epicratonic depositional setting may have limited wave and tidal energy, and the low gradient of the craton may have allowed for rapid movement of the shoreline over great distances. The pattern illustrated in these rocks may be applicable to other fine-grained and extensive deposits of epicratonic settings, where relatively passive transgression leads to widespread and uniform microstratigraphic successions.

CONCLUSIONS

A stratigraphically thin (< 10-m-thick) transgressive succession with uniform microstratigraphy exists in upper Famennian strata throughout much of the Rocky Mountains. It includes a thin basal, discontinuous sandstone bed, a thin shale bed, and an overlying oncolite-bearing carbonate wackestone with marine fauna. The lower shale bed is ~0.10 m thick and contains abundant spinicaudatans, which are known to inhabit very shallow, fresh to brackish water in shoreline and coastal plain environments. The presence of normal marine fauna with the spinicaudatans points to transitions in marine conditions associated with initial transgression. The areal extent of this bed of shallow-water fauna (~1600 north to south by ~1000 km east to west) represents one of the most remarkable paleoecological events of the latest Devonian in Laurentia. Moreover, this bed is highly unusual sedimentologically with respect to both its deposition and preservation. Nearshore and shoreline deposits typically show considerable facies heterogeneity and discontinuous stratal patterns resulting from irregular shoreline geometries, fluvial sediment point sources, and temporal and spatial variations in shallow-water processes, such as tides, waves, and currents. In addition, transgressive marginal marine and coastal deposits are often poorly preserved due to ravinement associated with shoreline retreat. The sedimentological and faunal data suggest remarkably passive and uniform transgression of a unique Late Devonian quiet-water shoreline system, and preservation of marginal marine fauna that tracked relative sea-level rise.

Sedimentological and paleontological evidence, including high organic content, pyritized fauna at the transition into the basal part of the overlying oncolite unit, diminutive fauna, and the prasinophyte algae *Tasmanites* at the base of the spinicaudatan bed, suggest temporary or seasonal hypoxia during deposition of the shale. The latter effectively use reduced nitrogen species, and large blooms at times lead to consumption of oxygen and eutrophication, and thus they are commonly a disaster taxon (Tappan 1980). Such processes commonly occur at the base of transgressive systems tracts (Prauss 2007) when nutrients were liberated from surface sediment or brought in with advancing water masses. By analogy with Early Jurassic epifaunal bivalves (Caswell and Coe 2013), the short-term expansion of the spinicaudatans may also reflect opportunistic behavior brought on by *Tasmanites* blooms and temporary hypoxia, which alternated with normal marine conditions and influx of marine fauna. The nature of spinicaudatan reproduction, including r-type selection, large larval broods, short lifespans, and small body sizes is consistent with stressed environments, including hypoxic conditions.

Integration of carbon and sulfur isotopic data with biostratigraphic and sedimentological analyses sheds light on the paleoenvironmental conditions during the deposition of the unique units of this study. The variations in the $\delta^{34}\text{S}$ data may reflect deposition in an epicontinental seaway with restricted circulation, locally brackish conditions, and possible temporary hypoxia. A $\sim 2\%$ negative shift in $\delta^{13}\text{C}_{\text{carb}}$ starting at the base of the oncolitic unit is a prelude to a long-term rise in isotopic values leading up to the base of the Mississippian. The unconformity at the base of the stratigraphic interval, with overlying basal Middle *expansa* Zone strata, likely reflects global regression and subsequent transgression, respectively, associated with the well-known Late Devonian (upper Lower *expansa* Zone) Dasberg event, but $\delta^{13}\text{C}_{\text{carb}}$ in strata below are too sparse to capture an isotopic signal. The thin stratigraphic interval of this study, with a record of passive transgression and remarkably widespread colonization by marginal marine fauna, resulted from unusual depositional processes and provides a window into distinctive paleoenvironmental conditions of the Late Devonian in western Laurentia. This case study provides a template for the interpretation of other epicratonic strata with uniform and widespread marginal marine fauna.

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SUPPLEMENTAL MATERIAL

Data are available from the PALAIOS Data Archive: <http://www.sepm.org/pages.aspx?pageid=332>.

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