Recognition and significance of Upper Devonian fluvial, estuarine, and mixed siliciclastic-carbonate nearshore marine facies in the San Juan Mountains (southwestern Colorado, USA): Multiple incised valleys backfilled by lowstand and transgressive systems tracts

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ABSTRACT

The Upper Devonian Ignacio Formation (as stratigraphically revised) comprises a transgressive, tide-dominated estuarine depositional system in the San Juan Mountains (Colorado, USA). The unit backfills at least three bedrock paleovalleys (10–30 km wide and ≥42 m deep) with a consistent stratigraphy of tidally influenced fluvial, bayhead-delta, central estuarine-basin, mixed tidal-flat, and estuarine-mouth tidal sandbar deposits. Paleovalleys were oriented northwest while longshore transport was to the north. The deposits represent Upper Devonian lowstand and transgressive systems tracts. The overlying Upper Devonian Elbert Formation (upper member) consists of geographically extensive tidal-flat deposits and is interpreted as mixed siliciclastic-carbonate bay-fill facies that represents an early highstand systems tract. Stratigraphic revision of the Ignacio Formation includes reassigning the basal conglomerate to the East Lime Creek Conglomerate, recognizing an unconformity separating these two units, and incorporating strata previously mapped as the McCracken Sandstone Member (Elbert Formation) into the Ignacio Formation. The Ignacio Formation was previously interpreted as Cambrian, but evidence that it is Devonian includes reexamined fossil data and detrital zircon U-Pb geochronology. The Ignacio Formation has a stratigraphic trend of detrital zircon ages shifting from a single ca. 1.7 Ga peak to bimodal ca. 1.4 Ga and ca. 1.7 Ga age peaks, which represents local source-area unroofing history. Specifically, the upper plate of a Proterozoic thrust system (ca. 1.7 Ga Twilight Gneiss) was eroded prior to exposure of the lower plate (ca. 1.4 Ga Uncom磅ahgre Formation). These results are a significant alternative interpretation of the geologic history of the southern Rocky Mountains.

INTRODUCTION

Over the past several decades there has been increasing recognition of the significance of incised valley fills (Vail et al., 1977; Van Waggoner et al., 1990; Allen and Posamentier, 1993; Catuneanu, 2006) and transgressive estuarine depositional systems ( Cotter and Driese, 1998; Fiaschein et al., 2009; Ainsworth et al., 2011) in evaluating relative sea-level changes and the influence of allogenic controlling variables (eustasy, tectonics, and sediment supply). In outcrop studies, the recognition of paleovalleys is complicated by available exposure versus the scale of the features. Similarly, estuarine facies may be difficult to recognize because of lateral variability and extent, compared again to available exposure. This study presents a new, integrated interpretation of the Upper Devonian sedimentary record for the southern Rocky Mountains based upon a depositional systems analysis of the Ignacio Formation.

There have been significant disagreements about the age, stratigraphy, and depositional environments of the Ignacio Formation in the San Juan Mountains of southwestern Colorado, USA (Fig. 1). The unit has been variously considered Cambrian or Devonian (Fig. 2), and depositional interpretations have ranged from shallow marine (Barnes, 1954; Baars, 1965; Baars and See, 1968) to colluvial fans, braided streams, lagoon-tidal flat, and marine shelf deposits (Wiggin, 1987) to estuarine and tidal-flat deposits (Maurer and Evans, 2011, 2013; McBride, 2016a). Key to understanding the geologic history of the study area are four significant modifications of existing stratigraphic relationships introduced in a companion paper (Evans and Holm-Denoma, 2018) and discussed further in this report.

First, an enigmatic conglomeratic unit locally overlying Proterozoic basement rocks and typically considered part of the overlying Ignacio Formation (Baars, 1966; Baars and See, 1968; Wiggin, 1987; Campbell and Gonzalez, 1996; Thomas, 2007) has been proposed as a new stratigraphic unit, the East Lime Creek Conglomerate (Evans and Holm-Denoma, 2018). The East Lime Creek Conglomerate is 0–23 m thick, consists of cobble-boulder conglomerate and thin interbedded sandstone, and has buttressing relationships to the underlying Proterozoic rocks interpreted as paleo-sea cliffs, paleo–wave-cut platforms, and paleo-tombolos. The unit has been interpreted as a rocky shoreline depositional system composed of upper shoreface-beachface tabular cobble-boulder gravels and offshore subaqueous debris-flow deposits (Evans
Figure 1. Location map showing the study area in southwestern Colorado (USA) and locations of measured sections (numbers refer to sections in Table 1 and Figs. 3 and 15). Locations without numbers were used for samples, paleocurrents, and additional observations but not measured sections. Regional geology is modified from Steven et al. (1977) and Evans and Reed (2007).
and Holm-Denoma, 2018). Because of poor age constraints, the age of the unit was previously considered Neoproterozoic–Cambrian (Wiggin, 1987; Campbell, 1994a, 1994b; Condon, 1995; Gonzales et al., 2004; Evans, 2007; McBride, 2016a), although Spoelhof (1976) and Wiggin (1987) proposed that it might be as young as Devonian. All of these previous age interpretations assumed that the overlying Ignacio Formation is Cambrian (see below). However, the unit was found to have a single late early Silurian (436 ± 17 Ma) detrital zircon, and this combined with field relations suggest that the unit is probably Lower Devonian (Evans and Holm-Denoma, 2018).

Second, there is a low-angle (<10°) disconformity between the East Lime Creek Conglomerate and Ignacio Formation, although it can be difficult to observe at some locations due to obscure bedding attitudes in the uppermost conglomerate beds (Evans and Holm-Denoma, 2018). In addition, sandstone in the East Lime Creek Conglomerate has sericite (sensu Eberl et al., 1987) cements, unlike the overlying Ignacio Formation. The sericite cements are interpreted as evidence of surficial weathering and early diagenetic alteration of primary mixed-layer illite and/or smectite clays to fine-grained muscovite and/or phengite prior to deposition of the overlying units. Finally, erosional reworking of the top of the unit is indicated by rare sericite-cemented sandstone clasts incorporated into the overlying Devonian units (Evans and Holm-Denoma, 2018).

Third, although many previous workers have considered the age of the Ignacio Formation to be Cambrian (Cross et al., 1905a, 1905b; Knight and Cooper, 1955; Baars and Knight, 1957; Rhodes and Fisher, 1957; Baars, 1966; Baars and See, 1968; Campbell 1994a, 1994b; Thomas, 2007), there is now strong evidence that the unit is Devonian. Briefly, the key new age considerations are that: (1) individual beds have been found containing the problematic Cambrian (?) Obulus brachiopods alongside Late Devonian placoderm fish fossils (McBride, 2016a); (2) the unit contains Cambrian and Ordovician detrital zircons (McBride, 2016a); and (3) the Ignacio Formation overlies the Lower (?) Devonian East Lime Creek Conglomerate (Evans and Holm-Denoma, 2018). In addition, some previous workers suggested the Ignacio Formation could be Devonian because of the absence of an unconformity between the Ignacio Formation and the overlying Devonian Elbert Formation (Read et al., 1949; Barnes, 1964; Maurer, 2012).
Finally, strata in the San Juan Mountains that were previously considered to be part of the overlying Devonian McCracken Sandstone Member of the Elbert Formation have been reassigned to the Ignacio Formation (Evans and Holm-Denoma, 2018). The McCracken Sandstone Member is defined from one exploration well in the Paradox Basin (the Four Corners area) (Cooper, 1955; Baars and Knight, 1957), and there have been significant disagreements about whether or not the unit is exposed in the San Juan Mountains. Baars (1965) argued that the unit did not extend far enough eastward to appear in the San Juan Mountains. Other workers have argued that the McCracken Sandstone Member can be recognized in the San Juan Mountains using the criteria that it is generally whiter in color, harder (due to silica cement), more quartzose rich, better sorted, and has better rounded grains than sandstone of the Ignacio Formation (Knight and Cooper, 1955; Baars and See, 1968; Campbell and Gonzales, 1996; Thomas, 2007; McBride, 2016a). However, those distinctions are not statistically robust—on petrofacies plots, there is significant overlap within one standard deviation (Evans and Holm-Denoma, 2018). McBride (2016a) proposed remapping the Ignacio Formation and McCracken Sandstone Member as two coeval, geographically adjacent units. In contrast, Evans and Holm-Denoma (2018) reassigned these strata to the Ignacio Formation because of the: (1) absence of an unconformity between the units; (2) absence of statistically significant petrologic differences; (3) evidence from facies analysis and paleocurrent data that the units formed an integrated depositional system (Maurer and Evans, 2011, 2013); and (4) geochronologic evidence that both units are Upper Devonian.

**METHODS**

Eighteen (18) stratigraphic sections were measured in the field and their locations recorded using GPS (Table 1). Sixty-six (66) thin sections for sandstone petrography were prepared using standard methods, and composition was determined from point counting >300 grains per slide (total n = 18,769) following the methods of Dickinson et al. (1983) (Table S1 in the Supplemental Materials†). Paleocurrent interpretations were based upon 124 unidirectional measurements of cross-bedding and from 24 bidirectional measurements from the orientation of the crests of wave ripple marks exposed on bedding surfaces. Vector means were calculated and plotted for each location, and rose diagrams were created using a nonlinear scale (Nemec, 1988), showing vector mean, circular standard deviation (Krause and Geijer, 1987), and Rayleigh test of significance (Curry, 1956). The paleocurrent data sets are statistically significant (p < 0.05). Five samples were collected for detrital zircon U-Pb geochronology. For each sample, 2 kg were disaggregated using standard crushing techniques, then zircons were concentrated using heavy liquid and magnetic separation. Zircon grains were mounted in epoxy, polished, and evaluated for zoning and i.

**TABLE 1. LOCATIONS OF STRATIGRAPHIC SECTIONS**

| Location number† | Informal description | Latitude | Longitude |
|-----------------|----------------------|----------|-----------|
| 1  | Fall Creek       | 37.50278°N | 107.5503°W |
| 2  | Canyon Creek     | 37.85111°N | 107.7314°W |
| 3  | Bakers Bridge    | 37.45889°N | 107.8006°W |
| 4  | Shalona Lake railroad outcrop | 37.48556°N | 107.8058°W |
| 5  | Rockwood         | 37.48694°N | 107.8078°W |
| 6  | Milepost 53.5 (U.S. Highway 550) | 37.67639°N | 107.7911°W |
| 7  | Milepost 54 (U.S. Highway 550) | 37.88056°N | 107.7861°W |
| 8  | Coal Bank Pass south | 37.69722°N | 107.7775°W |
| 9  | Meadow below Coal Bank Pass | 37.68556°N | 107.7606°W |
| 10 | West side of Lime Creek | 37.70889°N | 107.7592°W |
| 11 | Type section of the ELCC† | 37.70917°N | 107.7375°W |
| 12 | East side of Lime Creek | 37.70944°N | 107.7347°W |
| 13 | Andrews Lake trail | 37.71444°N | 107.7147°W |
| 14 | Molas Creek waterfall | 37.73944°N | 107.6681°W |
| 15 | East of Molas Lake | 37.73749°N | 107.6772°W |
| 16 | Sultan Creek–Molas Creek valley | 37.75751°N | 107.6576°W |
| 17 | Sultan Creek south | 37.71001°N | 107.6753°W |
| 18 | Sultan Creek north | 37.76501°N | 107.6753°W |

Note: Coordinates are in reference to North American Datum of 1927.
†Location numbers refer to Figures 1 and 3.
†ELCC—East Lime Creek Conglomerate (Evans and Holm-Denoma, 2018).

Research Laboratory in Denver, Colorado. Zircon was ablated with a Photon Machines Excite 193 nm ArF excimer laser in spot mode (150 total bursts per grain) with a repetition rate of 5 Hz, laser energy of −3 mJ, and an energy density of 4.11 J/cm². Pit depths were typically <20 µm. The rate of He carrier gas flow from the HeLex cell of the laser was ~0.6 L/min. Make-up Ar gas (~0.2 L/min) was added to the sample stream prior to its introduction into the plasma. Nitrogen with a flow rate of 5.5 mL/min was added to the sample stream to allow for significant reduction in ThO/Th (<0.5%) and improved ionization of refractory Th (Hu et al., 2008). Laser spot sizes for zircon were ~25 µm. With the magnet parked at a constant mass, the flat tops of the isotope peaks of 201Hg, 204Hg (Hg + Pb), 208Pb, 209Pb, 210Pb, 212Th, 232Th, and 235U were measured by rapidly deflecting the ion beam with a 30 s on-peak background measured prior to each 30 s analysis.

Raw data were reduced offline using the Iolite 2.5 program (Paton et al., 2011) to subtract on-peak background signals, correct for U-Pb downhole fractionation, and normalize the instrumental mass bias using external mineral reference materials, the ages of which had previously been determined by isotope dilution–thermal ionization mass spectrometry (ID-TIMS). Ages were corrected by standard sample bracketing with the primary zircon reference material Temora2 (417 Ma; Black et al., 2004) and secondary reference material Plešovice (337 Ma; Sláma et al., 2008) and USGS standard WRP-63-08 (1707 Ma). Reduced data were compiled into a probability density diagram using Isoplot 4.15 (Ludwig, 2012). 206Pb/238U ages are reported for zircon...
analyses <ca. 1300 Ma, and 207Pb/206Pb ages are used for older ages (Gehrels, 2012). Analyses with discordance >20% were excluded from probability density plots (Table S2 [footnote 1]).

## RESULTS

### Stratigraphic Relationships

#### Description

The stratigraphic sections are correlated using the conformable contact between the Ignacio Formation (as redefined) and the overlying upper member of the Elbert Formation (Fig. 3). This upper contact marks a change from the dominantly fluvial and estuarine sandstone of the Ignacio Formation to the marine shale and carbonate of the upper member of the Elbert Formation. Outcrop correlation reveals three important relationships.

First, the Ignacio Formation is lithologically heterogeneous (sandstone, mudrocks, and minor carbonate, conglomerate, and replaced evaporite) and stratigraphically complex, with different lithologies dominant at intervals throughout the section. A number of locations have a sandstone-rich interval at the top of the section, which was the basis for previous workers calling this interval the McCracken Sandstone Member. However, sandstone-rich intervals are interspersed throughout the Ignacio section, and the uppermost part of the unit is not sandstone rich at all locations.

Second, the thickness of the Ignacio Formation is highly variable between locations. Where exposed, the basal contact is an unconformity that rests on either Proterozoic crystalline rocks or the East Lime Creek Conglomerate, with erosional relief typically <2.5 m. Overall, the Ignacio Formation varies locally between 0 and 42 m thick.

Third, the relationship of Devonian sedimentary rocks to underlying Proterozoic basement shows the importance of paleotopography (Fig. 3). Using the measured sections and observations at additional locations where the unit is not sandstone rich at all locations.

#### Interpretation

Previous workers advanced two arguments to explain lithologic variation and complex stratigraphic relations in the lower Paleozoic section in the San Juan Mountains. The first argument is that syndepositional faulting controlled Ignacio Formation deposition and explains the juxtaposition of the Ignacio Formation and Proterozoic rocks (Baars, 1966; Baars and See; 1968; Thomas, 2007). The second argument is that the local presence or absence of the conglomerate, Ignacio Formation, McCracken Sandstone Member, upper member of the Elbert Formation, and/or combinations of those units was due to a complex history of deposition, pre-Pennsylvanian block uplifts (Grenadier horst, Sneffels horst, and unnamed horsts and grabens), erosion, subsidence, and further deposition (Baars, 1966; Baars and See; 1968; Weimer, 1980; Baars et al., 1987); according to this argument, stratigraphic relationships required recurrent movement on faults, and even reversals of movement sense on individual faults.

Most of the evidence for the first proposal is from what is now recognized as the East Lime Creek Conglomerate rather than the Ignacio Formation. Detailed examination reveals that neither unit shows (1) evidence for fault controlled proximal-distal trends in lithology, thickness, or grain size; (2) tectonically controlled changes in facies distributions; (3) changes or reversals in paleocurrent patterns; or (4) changes in sediment source areas. Further inspection of the hypothesized fault zones failed to show any direct evidence for faults (i.e., drag folds, slickensides, or fault gouge), nor evidence for seismogenic features in the adjacent sedimentary units, such as fluid-escape structures or convoluted bedding (e.g., Evans, 1994; Myrow and Chen, 2015).

Misinterpretation of different depositional facies within the Ignacio Formation as different stratigraphic units is responsible for the second proposal. As a laterally extensive depositional system, the Ignacio Formation presents different depositional facies at different locations. The presence or absence of depositional facies at any location does not require faulting. It should be noted that McBride (2016a) used a different stratigraphic approach yet reached substantially the same conclusion about faulting. In summary, this study finds no evidence for post-Proterozoic but pre-Pennsylvanian (i.e., pre–Ancestral Rocky Mountains) faulting in the study area.

### Facies Analysis

Twenty-six (26) lithofacies are observed in the Ignacio Formation (Table 2). Following standard practice, these lithofacies are grouped into seven lithofacies associations (FA1–FA7) representing specific depositional environments or subenvironments (Table 3).

#### Facies Association 1 (FA1): Fluvial Deposits

**Description.** FA1 consists of channeliform sandstone bodies interpreted as single-story and multistory channel fills, and intervening fine-grained deposits with poorly developed paleosols (Fig. 4). The lower parts of channel-fill successions consist of red trough cross-bedded sandstone and planar-tabular cross-bedded sandstone that form sets up to 50 cm thick and stacked cosets up to 1.5 m thick (Figs. 5A, 5B). Cross-bed sets are separated by internal erosion surfaces and channel lags consisting of thin pebble conglomerate layers (Fig. 5B), rare mudstone intraclasts, and massive pebbly sandstone. The upper parts of channel-fill successions include ripple-laminated sandstone
Figure 3. Measured stratigraphic sections in the Ignacio Formation. Numbers refer to locations shown in Figure 1. Coordinates are given in Table 1. Horizontal correlation line marks the contact with the upper member of the Elbert Formation. MP—highway milepost; ELCC—East Lime Creek Conglomerate; mbr.—member; Fm.—Formation. Grain-size abbreviations: ms—mudstone; vfg—very fine-grained sandstone; fg—fine-grained sandstone; mg—medium-grained sandstone; cg—coarse-grained sandstone; vcg—very coarse-grained sandstone; pbl—pebble conglomerate; cb—cobble conglomerate; b—boulder conglomerate. (Continued on following page.)
Symbols:
- conglomerate, intraclasts
- planar-tabular cross-bedding
- trough cross-bedding
- planar lamination
- current ripples
- climbing ripple lamination
- wave ripples
- flaser, wavy, or lenticular bedding
- mudcracks
- mudcrack breccias
- burrows
- surface traces
- shells
- bioherm

Grain-size scale

Figure 3 (continued).
### TABLE 2. LITHOFACIES CODES AND DESCRIPTIONS

| Code | Lithology* | Sedimentary structures | Interpretation |
|------|------------|------------------------|----------------|
| Gm   | Conglomerate, pebble | Massive | Thin fluvial or estuarine channel lags |
| Smc  | Sandstone, cg–vcg | Massive, can be pebbly | Rapid deposition, destratified, lags |
| Smn  | Sandstone, fg–cg | Massive, normal grading | Fallout from suspension |
| Smf  | Sandstone, vfg–mg | Massive | Rapid deposition or destratified |
| Sp   | Sandstone, mg–vcg | Planar-tabular cross-bedded | Two-dimensional dunes |
| Sph  | Sandstone, mg–cg | Herringbone cross-bedded | Two-dimensional dunes with flow reversals |
| St   | Sandstone, mg–vog | Trough cross bedded, festoon cross-bedded | Three-dimensional dunes |
| Sr   | Sandstone, vfg–mg | Asymmetrical ripples, climbing ripples | Current ripples |
| Sw   | Sandstone, vfg–mg | Continuously crested symmetrical ripples | Wave ripples |
| SI   | Sandstone, vfg–lg | Low-angle, planar to wavy laminated | Low-angle inclined planar bedding |
| Se   | Sandstone, vfg–vcg | Massive with mud intraclasts | Mud clasts derived from mudcracks |
| SSI  | Siltstone | Planar laminated | Fallout from suspension |
| SSm  | Siltstone | Massive | Rapid deposition or destratified |
| SMf  | Heterolithic (ss–ms) | Flaser bedded | Ripples with mud drapes |
| SMw  | Heterolithic (ss–ms) | Wavy bedded | Ripples with mud drapes |
| SMk  | Heterolithic (ss–ms) | Lenticular bedded | Ripples with mud drapes |
| SMl  | Heterolithic (ss–ms) | Planar laminated | Tidalites |
| Fl   | Mud shale | Planar laminated | Fallout from suspension setting |
| Fm   | Mudstone | Massive | Rapid deposition or destratified |
| Cm   | Carbonate, mudstone | Massive | Carbonate mud (micrite) |
| Cp   | Carbonate, packstone | Massive with shell hash | Storm layer with shell debris |
| Cf   | Carbonate, floatstone | Massive with mud clasts | Mud clasts derived from mudcracks |
| Cb   | Carbonate, bindstone | Laminated (planar, wavy, or domal) | Biostratification features |
| Cn   | Carbonate, nodular | Disrupted bedding, teepees, hoppers, nodules | Desiccation or evaporite dissolution |
| P    | Pedogenic carbonate | Small nodules | Poorly developed paleosol |
| K    | Coal | Coal fragments | Wood debris |

*Abbreviations: vcg—very coarse grained; cg—coarse grained; mg—medium grained; fg—fine grained; vfg—very fine grained; ss—sandstone; ms—mudstone.

### TABLE 3. FACIES ASSOCIATIONS

| Code | Facies association | Lithofacies | Ichnofacies | Organization | Interpretation |
|------|-------------------|-------------|-------------|--------------|----------------|
| FA1  | Fluvial channel and floodplain | St, Sp, Sr, Gm, Smc, St, Smf, SStm, Fm | Rare (Sk) | Broadly lenticular, erosive base and internal erosion surfaces, 3–6 m thick, mostly sandy two- and three-dimensional dunes | Multistory channel fill |
| FA2  | Tidally influenced fluvial channel | St, Sp, Smc, SPh, Srw, SMf, SMw, SMk, FI, Fm | Rare (Sk, P) | Broadly lenticular, erosive base, flow reversals, drapes, mud chips | Estuarine point bar |
| FA3  | Bayhead delta | St, Sp, Sr, Gm, Sm, SI, SMf, SMw, SMk, FI, Fm | Minor (Sk) | Inclined heterolithic strata, climbing ripples, wave-modified turbidites | Bayhead delta |
| FA4  | Estuarine channel and central basin | St, Sp, Sr, Srw, SI, Smt, SMf, SMw, SMk, FI, Fm | Major: P, Pa, Th, RO, DI, MO, TR, GA | Broadly lenticular sand bodies with mud drapes, extensive bioturbation | Central estuarine basin (local firmgrounds) |
| FA5  | Tidal flat (mostly siliciclastic) | Sli, Srw, Smf, Se, SSI, SMf, SMw, SMk, SMI, Fm | Major: Ru, SK, P | Sheet like, poorly exposed, with small channels, mud drapes, mud chips | Estuary margin |
| FA6  | Peritidal and supratidal flat (mostly carbonate) | Cm, Cp, CIf, Cb, CN, Fm | Biostratification | Small bioherms, disrupted bedding, teepee structures, nodules, hoppers | Estuary margin |
| FA7  | Estuarine-mouth tidal | St, Sph, Srw, St, Smf, Fm | Major: P, Th, LO | Tabular, internal erosion surfaces, may be capped by firmgrounds | Tidal sand bar |

Note: See Table 2 for explanation of lithofacies codes. Ichnofacies codes: Di—Diplocraterion; Ga—Gastrochaenolites; Lo—Lockea; Mo—Monomorphichnus; P—Planolites; Pa—Palaeophycus; Ro—Rosselia; Ru—Rusophycus; Sk—Skolithos; Th—Thalassinoides; Tr—Trichophycus.
Canyon Creek

US Highway 550 MP 53.5

Bakers Bridge

Symbols:

- conglomerate, intraclasts
- planar-tabular cross-bedding
- mudcracks
- rough cross-bedding
- mudcrack breccia
- planar lamination
- burrows
- current ripples
- surface traces
- climbing ripple lamination
- shells
- wave ripples
- bioherm

Figure 4. Detailed stratigraphic sections at Canyon Creek (location 2 in Figs. 1 and 3), along U.S. Highway 550 at milepost (MP) 53.5 (location 6), and at Bakers Bridge (location 3). Lithofacies (and their codes) are described in Table 2. Facies associations (bold italics) are described in Table 3. Abbreviations: FS—flooding surface; F/TRS—fluvial/tidal ravinement surface; MFS—maximum flooding surface; SB—sequence boundary; TSE—transgressive surface of erosion; Fm.—Formation; IHS—inclined heterolithic stratification. See Figure 3 for definitions of grain-size abbreviations and lithology symbols.
Figure 5. Outcrop photographs of fluvial (facies association FA1; see Table 3) and tidally-influenced fluvial (FA2) deposits. (A) Thin conglomerate (lithofacies Gm; see Table 2) at the base of fluvial channel-fill with imbricated clasts. Scale bar in centimeters. (B) Cosets of trough cross-bedded sandstone (lithofacies St) and overlying granule conglomerate (lithofacies Gm). Arrows indicate set boundaries. Scale bar is 15 cm. (C) Proximal overbank sequence of stacked thin, massive, fine-grained red sandstones (lithofacies Smf) and thin, laminated, coarse-grained white sandstones (lithofacies Sl). Subsequent fluid flow through the better-sorted, coarser layers bleached the iron oxides. Scale bar is 15 cm. (D) Single sets of trough cross-bedded sandstone (lithofacies St) and ripple-laminated sandstone (lithofacies Sr) with mud drapes (lithofacies Fm). Hammer is 30 cm. (E) Channel-fill sandstone with wedge-shaped beds of trough cross-bedded sandstone (lithofacies St) and mud drapes, overlying floodplain mudstones (lithofacies Fm). Hammer is 28 cm. (F) Thinning- and fining-upward (represented by white triangle) channel fill consisting of packages of massive fine-grained sandstones (lithofacies Smf), laminated sandstones (lithofacies Sl), and mud drapes (lithofacies Fm), overlying floodplain mudstones (lithofacies Fl). Hammer is 28 cm.
The interbedded finer-grained intervals are interpreted as proximal overbank vals of massive red fine-grained sandstone, siltstone, and mudstone (Fig. 5C). Locally, these are interbedded with planar laminated red sandstone, siltstone, and mud shale. The red finer-grained deposits are commonly interbedded with thin stringers of coarse-grained white sandstone (Fig. 5C). The white color probably represents bleaching of hematite during late diagenetic fluid migration through the coarser-grained and better-sorted layers.

Biotaurbation is generally absent, but there are rare examples of Skolithos in the upper parts of complete channel-fill successions. Finally, the upper parts of finer-grained successions show a combination of carbonate content and minor stratal disruption that suggests poorly developed calcsols.

Interpretation. FA1 is interpreted as the deposits of sand bedload to mixed-load fluvial systems that formed broadly lenticular channels with typical channel depths <7 m. Based on paleocurrent data, the primary channel-filling elements were downstream-accreting two- and three-dimensional dunes with intervening pool fills (thin conglomerate stringers and wedge-shaped cross-bedded sets in sandstone). Lateral accretion surfaces were not observed. The interbedded finer-grained intervals are interpreted as proximal overbank deposits. The presence of overbank deposits, rare mudstone intraclasts, and local coal fragments suggests some level of incipient bank stability and episodic bank failures (e.g., Plint, 1986). However, the interpreted floodplain deposits were dominantly noncohesive (sand- and silt-sized) materials. This implies that the banks of paleochannels were relatively nonresistant, low, and easily overtopped, which is supported by the observed poorly developed paleosols. Finally, although Skolithos is typically found in marine rocks, it is known to occur in fluvial deposits (e.g., Trewin and McNamara, 1994).

Facies Association 2 (FA2): Tidally Influenced Fluvial Deposits

Description. FA2 deposits resemble those of FA1 as single and multistory channel-fill successions dominated by trough cross-bedded and planar-tabular cross-bedded sandstone units separated by internal erosion surfaces with interbedded pebble stringers. As with FA1, the deposits are organized into broadly lenticular channel bodies up to ~5 m thick. The key difference between FA1 and FA2 is evidence for current reversals in FA2, including: (1) multistory channel deposits with interbedded herringbone cross-bedded sandstone; (2) thin interbedded intervals of heterolithic flaser-bedded, wavy-bedded, or lenticular-bedded sandstone-mudstone couplets; and (3) thin mudstone drapes between cross-bed sets (Fig. 5D). Partially mud-draped cross-bed surfaces suggest remolding by tidally influenced flow modification (Fig. 5E). Fining-and thinning-upward channel-fill sequences are observed (Fig. 5F) and may represent channel abandonment and avulsion processes. Rare trace fossils (Skolithos and Planolites), mud-chip intraclasts, and wave-rippled sandstone are also present in FA2.

Interpretation. FA2 is interpreted as tidally influenced fluvial deposits because of the combination of similar appearance, organization, and architecture to FA1, and the presence of heterolithic flaser, wavy, and lenticular bedding (Reineck and Wunderlich, 1968); evidence for flow reversal (herringbone cross-bedding); wave ripples; and trace fossils that may indicate brackish-water conditions (MacEachern et al., 2005). Tidal effects in modern estuaries are known to extend upstream into rivers for tens of kilometers (Allen, 1991).

Facies Association 3 (FA3): Bayhead Delta Deposits

Description. FA3 deposits are present at several locations in the study area. Key diagnostic features in these deposits are composite sets of meter-scale, low-angle sigmoidal trough cross-bedding with intervening complete to partial mud drapes (Fig. 6A). Similar deposits have been called inclined heterolithic stratification (IHS) (Thomas et al., 1987), and modern examples show the interaction of fluvial currents and tidal reversals (Fenies et al., 1999). Steep-sided, small-scale channeling is also notable in FA3 (Fig. 6B), demonstrating greater sediment cohesion in the finer-grained estuarine sediments versus sandy fluvial sediment in FA1 and FA2.

Repeated facies successions in FA3 include (1) vertically stacked packages of low-angle to vertically climbing ripple-laminated, fine- to medium-grained red sandstone (Fig. 6C); (2) packages consisting of (in ascending order) an erosion surface; massive, normally graded sandstone; planar-laminated sandstone; and capping wave ripples (Fig. 6C); and (3) intervals of interbedded herringbone cross-bedded sandstone, heterolithic flaser-bedded sandstone-mudstone couplets, laminated siltstone, and thin mudstone drapes. Locally interbedded with these facies successions are thin pebble stringers or lenses of pebbly coarse-grained sandstone (Figs. 6C, 6D). The upper surface of FA3 deposits is typically an erosional lag deposit (Fig. 4), and the contact between FA3 deposits and overlying estuarine central basin deposits (FA4) is marked by a color change from red to green-gray (Fig. 6E).

Interpretation. FA3 is interpreted as representing bayhead delta depositional environments, based upon diagnostic features and structures seen in modern and ancient examples (Aschoff et al., 2018; Simms et al., 2018). The low-angle, sigmoidal IHS is a common feature in bayhead deltas, representing river flows into mixed-salinity receiving water bodies. Intervals of climbing-ripple sandstone are interpreted as delta-foreset or distributary mouth-bar deposits caused by flow deceleration and rapid deposition of bedload (Wright, 1977). The event layers with wave-rippled upper surfaces are interpreted as wave-modified turbidites (Myrow et al., 2002). The presence of wave-modified turbidites in these deposits is consistent with episodic density underflows at the delta front into relatively shallow water affected by wave processes. Turbidity currents are the primary mode of sediment transport from fluvial channels to the delta front in bayhead deltas (Aschoff et al., 2018). The heterolithic deposits with tidal sedimentary structures are interpreted as tidalites (Longhitano et al., 2012): Intervals of tidalites interbedded with bayhead...
Figure 6. Field photographs of bayhead delta (facies association FA3; see Table 3) deposits. (A) Composite set of inclined heterolithic strata (IHS) showing sigmoidal trough cross-bedded sandstone (lithofacies St; see Table 2) and mud drapes (lithofacies Fm). Hammer is 28 cm. (B) Steep-sided channel incised into estuarine central basin deposits and infilled with massive fine-grained sandstones (lithofacies Smf) and mud drapes (lithofacies Fm). Scale bar is 15 cm. (C) Climbing ripple-laminated sandstone (lithofacies Sr), eroded and overlain by pebbly massive coarse-grained sandstone (lithofacies Smc) representing a discontinuously bedded, winnowed surface (fluvial-tidal ravinement surface). These are overlain by normally graded massive sandstone (lithofacies Smn), laminated sandstone (lithofacies Sl), and wave-rippled sandstone (lithofacies Srw), interpreted as a wave-modified turbidite. Scale bar is 15 cm. (D) Interbedded reddish massive fine-grained sandstones (lithofacies Smf) and white pebbly massive coarse-grained sandstones (lithofacies Smc) in the delta front. The erosional lags are interpreted as fluvial-tidal ravinement surfaces (arrows). Hammer is 28 cm. (E) Color transition from reddish bayhead delta (BHD) to gray-green central estuarine basin (CEB) and estuarine bar (EB) deposits (person for scale).
delta-front deposits probably represent progressive infilling of abandoned distributary channels.

Deposits of these episodic depositional events commonly overlie pebble stringers interpreted as erosional lags or bypass surfaces, representing fluvial-tidal reworking (Frey et al., 1989). The distinctive erosional surface and lag deposit at the top of FA3 is interpreted as a fluvial-tidal ravinement surface.

The distinctive red to gray-green color change at the bayhead delta–estuarine central basin transition (Fig. 6E) has been observed from other modern and ancient estuaries (Cotter and Driese, 1998; Boyd, 2010). Reddening in the Ignacio Formation is facies specific, indicating that it is not due to diageneric reddening of the unit as a whole. There is no evidence that reddening is related to changes in lithology or permeability. We suggest that the reason why red colors are specific to the fluvial and deltaic portions of this unit is probably related to the effect of terrestrial weathering in the source areas. In other words, the color transition from fluvial and bayhead delta sediments to estuarine sediments is due to significant input of red sand and mud into the proximal reaches of the estuarine environment, such as observed in the modern Bay of Fundy estuary (southeastern Canada).

Many of the described features are characteristic of marine deltas in general, such as basinward-directed paleocurrents; mixture of fluvial, wave, and tidal features; trace fossils and fauna indicative of mixed-salinity conditions; and evidence for shallow water depths (Aschoff et al., 2018; Simms et al., 2018). In this study, the bayhead delta interpretation is additionally based upon the relatively small scale of the features (<10 m thick), close association with underlying tidally influenced fluvial deposits, the capping erosion surface interpreted as a fluvial-tidal ravinement surface (e.g., Simms et al., 2018), and the overlying estuarine central basin deposits (FA4).

**Facies Association 4 (FA4): Estuarine Channel and Central Basin Deposits**

**Description.** FA4 consists of finer grained, variegated gray-green to red, heterolithic sandstone and shale, with fossils and trace fossils indicative of mixed-salinity conditions. The most prevalent deposits are thin, laterally continuous sandstone-mudstone couplets with flaser, wavy, or lenticular bedding (Fig. 7A). These sheet-like deposits are locally incised by broadly lenticular channel fills up to ~2 m thick. These channel-fill successions are either single-story trough cross-bedded sandstones encased in finer-grained sediment (Fig. 7B) or multistory trough cross-bedded sandstones consisting of single cross-bed sets 30–50 cm thick overlain by thin mud drapes (Fig. 7C). Most bedding shows some degree of biological mixing, homogenization, and loss of primary sedimentary structures (Fig. 7C). In addition, certain horizons are intensely bioturbated (Fig. 7D).

Trace fossils observed in FA4 include *Planolites*, *PalaepHyicus*, *Thalassinoides*, *Rosselia*, *Monomorophichnus*, *Diplocraterion*, *Skolithos*, *Gastrochaenolites*, and *Trichophycus* (Figs. 7D–7G). Most individual bedding surfaces display low-diversity trace-fossil assemblages dominated by *Planolites* or dominated by a combination of *Rosselia* and *Thalassinoides*. Vertical mixing extended downward ~30 cm in some instances, resulting in beds with indistinct burrow motting and loss of primary features. Most of the trace fossils listed above were also observed by Wiggins (1987) and McBride (2016a), although Wiggins (1987) also recorded *Arenicolitites* and *Corophoides*.

Most fossils in the Ignacio Formation (including the reassigned strata from the McCracken Sandstone Member) are found in FA4. Vertebrate fossils consist entirely of placoderm fish plates. Previous workers have described *Bothriolepis coloradensis*, *Bothriolepis canadenis*, *Bothriolepis major*, *Bothriolepis leidyi*, *Holoptychius giganteus*, and *Holoptychius tuberculatus* (Eastman, 1904; Cross and Larsen, 1935; Denison, 1951), although Thomson and Thomas (2001) argued that *B. coloradensis* and *B. leidyi* cannot be distinguished from *Bothriolepis nitida*. These all have late Frasnian–Famennian faunal ages (Thomson and Thomas, 2001). Other fossils include poorly preserved, phosphatic brachiopods equivocally identified as *Obulus* sp. (Cross et al., 1905a; Rhodes and Fisher, 1957), *Lingulella sp.*, and *Dicelomus sp.* (Baars, 1965). The poor quality of brachiopod samples has precluded better taxonomic determinations and limited their usefulness for age determinations (Read et al., 1949; Barnes, 1954; Baars and Knight, 1957; Wiggins, 1987; McBride, 2016a). Although *Obulus* has been considered late Cambrian–Ordovician in age (Emig, 2002), in the study area these *Obulus* (?) brachiopods have been recovered from the same bedding unit as Upper Devonian placoderm fish plates (McBride, 2016a).

**Interpretation.** The basis for interpreting FA4 as estuarine channel and central basin deposits includes the finer grain size (cf. FA1-3), variegated colors, prevalence of tidal bedding structures, architecture of estuarine channel fills, prevalence of bioturbation, and fossils consistent with brackish-marine conditions. The laterally continuous sandstone-mudstone couplets with flaser, wavy, or lenticular bedding are interpreted as tidalites (Reineck and Singer, 1980). The intensely bioturbated surfaces are interpreted as firmgrounds (Ekdale et al., 1984). These omission surfaces have been recognized as components of ravinement surfaces in other studies (Jordan et al., 2018).

These successions are similar to modern estuarine channel and central basin deposits observed in Willapa Bay (Washington State, USA) where successions consist of (1) bioturbated basal lags, overlain by (2) gently dipping interlaminated sand and mud layers of the accretionary bank, then overlain by (3) mudflat and supratidal flat deposits (Clifton and Phillips, 1980). In the Ignacio Formation, FA4 is dominated by tidalites interbedded with mudstone-rich intervals. Wave influence is indicated by wave ripples and by omission surfaces that acted as firmgrounds for benthic communities (Ekwenye et al., 2018). Brachiopod shell lag horizons are probable evidence for wave reworking (Frey et al., 1989). The trace fossils are consistent with those of modern and ancient estuarine systems (Howard and Fry, 1973; Hubbard et al., 2004) and represent a depauperate mixed *Skolithos*-Cruziana ichnofacies as observed in other estuarine environments (Ekdale et al., 1984; Hubbard et al., 2004). Both *Bothriolepis* and *Holoptychius* placoderm fish fossils have been found in Upper Devonian freshwater, estuarine, and coastal deposits. It has been speculated that these
Figure 7. Field photographs of estuarine channel and estuarine central basin deposits (facies association FA4; see Table 3). (A) Estuarine central basin deposits of heterolithic flaser-bedded (lithofacies SMf; see Table 2) and wavy-bedded tidalites (lithofacies SMw). Circled pen is 14 cm. (B) Small sand dune (lithofacies Srw) and mud drapes (lithofacies Fm) in estuarine central basin deposits. Scale bar is 15 cm. (C) Estuarine channel fill consisting of stacked trough cross-bedded sandstone (lithofacies St), ripple-laminated sandstone (lithofacies Sr and Srw), and mud drapes (lithofacies Fm). The lower sandstone has been disrupted and mottled by intense bioturbation (lithofacies Smf). Hammer is 28 cm. F/TRS—fluvial-tidal ravinement surface. (D) Firmground dominated by Planolites (P). Coin is 1.8 cm. (E) Firmground dominated by Thalassinoideas (Th), Rosselia (Ro), and Gastrochaenolites (Ga). Scale bar is 15 cm. (F) Firmground dominated by Monomorphichnus (Mo). Scale bar is 15 cm. (G) Firmground dominated by Diplocraterion (Di). Coin is 1.8 cm.
FA5 is interpreted as siliciclastic-dominated tidal-flat deposits on 15 November 2019.

Facies Association 5 (FA5): Siliciclastic Tidal-Flat Deposits

Description. FA5 deposits are fining-upward successions typically 1–2 m thick (Fig. 8A). The base of the succession is transitional to the estuarine channel fills discussed above. These are overlain by (in ascending order): (1) heterolithic, lenticular-bedded sandstone-mudstone couplets (Fig. 8B); (2) planar-laminated sandstone-mudstone couplets or siltstone-mudstone couplets (Fig. 8C); and (3) laminated siltstone, mud shale, and massive mudstone (Fig. 8D). The planar-laminated couplets show rare examples of double mud drapes. The upper mudstone commonly contains mudstone intraclasts (Fig. 8D), and locally bedding surface horizons are covered with mudstone intraclast accumulations (Fig. 8E). Trace fossils are patchy but locally abundant, including Rusophycus (Fig. 8F), Palaeophycus (Fig. 8G), Planolites, and Skolithos (Fig. 8H). There is a notable size reduction in the dimensions of Planolites and Skolithos compared with burrows in the estuarine central basin. Finally, FA5 successions are capped by carbonate-evaporite intervals of FA6 (see next section).

Interpretation. FA5 is interpreted as siliciclastic-dominated tidal-flat deposits. These are found either as lateral facies equivalents to the estuarine channels and central basin deposits in FA4, suggesting that they are similar to accretionary bank deposits (Clifton and Phillips, 1980), or overlying the estuarine channels of FA4, suggesting progradation of the tidal-flat environments (Frey and Howard, 1986; Daidu, 2013). Typically, FA5 transitions upward from subtidal estuarine channel fills and sandflat deposits with flaser-bedded or wavy-bedded sandstone and mudstone drapes, to intertidal mudflat deposits with sand-starved ripples and tidal rhythmites, to supratidal mudflat deposits (Fig. 8D). The planar-laminated couplets show rare examples of double mud drapes. The upper mudstone commonly contains mudstone intraclasts (Fig. 8D), and locally bedding surface horizons are covered with mudstone intraclast accumulations (Fig. 8E). Trace fossils are patchy but locally abundant, including Rusophycus (Fig. 8F), Palaeophycus (Fig. 8G), Planolites, and Skolithos (Fig. 8H). There is a notable size reduction in the dimensions of Planolites and Skolithos compared with burrows in the estuarine central basin. Finally, FA5 successions are capped by carbonate-evaporite intervals of FA6 (see next section).

Facies Association 6 (FA6): Carbonate Supratidal-Flat Deposits

Description. The uppermost part of tidal-flat successions in the Ignacio Formation are carbonate-evaporite dominated, and show significant signs of stratal disruption indicative of exposure, early cementation, evaporite dissolution, and replacement. The transition upward from FA5 deposits to FA6 deposits is gradational, marked by carbonate mudstone and rare shelly packstone interbedded with fine-grained siliciclastic red mudstone. Most carbonate mudstone contains small amounts (<5%) of silt- and sand-sized quartz, some of probable eolian origin (McBride, 2016a). The lower portions of FA6 deposits include meter-scale, flat-bottomed, domal carbonate accumulations that are internally massive or have cryptic lamination (Fig. 9A); relatively flat to wrinkled or pustular very fine-scale lamination (Fig. 9B); carbonate mudstone breccias (Fig. 9B); and replaced hopper crystals (Fig. 9C). In contrast, the upper portion of FA6 deposits includes horizons of disrupted nodular carbonates and carbonate breccia (Fig. 9E); poorly developed and truncated syndepositional folds in isolated carbonate layers (Fig. 9F); and horizons with upward-tilting carbonate layers thrust over adjacent layers (Fig. 9G). In parts of the study area, supratidal-flat sequences were dolomitized, possibly by Cenozoic hydrothermal alteration (Onasch et al., 1994; McBride, 2016b), to produce resistant beds that also contain thin silicified horizons (Fig. 9D).

Interpretation. FA6 is interpreted as indicating a peritidal carbonate depositional environment representing the upper intertidal to supratidal zone. The small domal structures are onlapped by adjacent strata (Fig. 9A), implying topographic relief in the depositional environment. The mostly massive or indistinct internal features imply that the domes were either thrombolite (Aitken, 1967), sponge, or microbial mounds in the upper part of the intertidal zone. The overlying flat or wrinkled or pustular finely laminated layers (Fig. 9B) are interpreted as microbial laminates produced by the trapping and binding of sediment by cyanobacteria biofilms (Logan et al., 1964). The carbonate mudstone intraclasts and mudstone breccias (Fig. 9B) are interpreted as accumulations of mud chips derived from mudcracked polygons. The tilted and overthrust carbonate horizons (Fig. 9G) are interpreted as teepee structures, which typically form from shrinking-swelling cycles in partly lithified carbonate having surface crusts or beachrock (Pratt, 2010), or from wrinkled microbial mats (Kendall, 2010).

There are multiple features indicative of dissolved or replaced evaporites in FA6, such as hopper crystal casts (Fig. 9C). The continuous horizons of disrupted carbonate nodules and breccias (Fig. 9E) have been observed elsewhere and interpreted as microkarst-collapse features produced due to the dissolution of underlying relatively thin layers of stratiform evaporite crystals (Middleton, 1961; Warren et al., 1990; Smith, 1997). Stratal disruption and syndepositional folding in individual carbonate horizons are suggestive of replaced evaporites that had enterolithic folding due to displacive crystal growth as part of the transition between gypsum and anhydrite (Kendall, 2010). Silification in the form of replacement chert produces blebs and anastomosing irregular layers at numerous localities (Fig. 9D), probably representing replacement of primary evaporite minerals (Milliken, 1979).

Overall, FA6 is dominated by laterally continuous carbonate mudstones that are mostly either internally massive, or microbial laminites, or disrupted layers with mud-chip breccias, mudcracks, nodules, teepee structures, and
Figure 8. Field photographs of siliciclastic tidal-flat deposits (facies association FA5; see Table 3). (A) Typical appearance of 1–2-m-thick prograding tidal flats, including a subtidal (ST) channel-fill sequence of massive and planar-laminated sandstone overlain by intertidal (IT) heterolithic sandstones and mudstones, overlain by supratidal (SuT) carbonates. Scale bar is 1 m. (B) Lenticular-bedded tidalites (lithofacies SMk; see Table 2) overlying mudstone (lithofacies Fl). Hammer is 28 cm. (C) Bundles (brackets) of tidal rhythms (lithofacies SMl) separated by thicker intervals of mudstone (lithofacies Fm). Hammer is 28 cm. (D) Uppermost intertidal sequence showing red mudstones (lithofacies Fm) with mudstone intraclasts (arrow; lithofacies Se) below carbonates of supratidal flat (lithofacies Cm). Hammer is 28 cm. (E) Bedding-plane surface showing mudstone intraclasts (lithofacies Se), interpreted as mud chips from intertidal zone mudcracks. Coin is 2.5 cm. (F) Firmground with Rusophycus (Ru) tracks. Coin is 1.8 cm. (G) Firmground dominated by *Palaeophycus* (Pa). Scale bar is 4 cm. (H) Firmground dominated by *Planolites* (P) and *Skolithos* (Sk). Scale bar is 15 cm.
Figure 9. Field photographs of carbonate peritidal (upper intertidal and supratidal zone) deposits (facies association FA6; see Table 3). (A) Small thrombolite or sponge or microbial bioclast [SB] overlying heterolithic fine-grained clastic deposits of the intertidal zone [IT] and overlain by carbonates in the supratidal zone [SuT]. Scale bar is 1.5 m. (B) Laminated microbialite [lithofacies Cb; see Table 2] overlying intraclast floatstone [lithofacies Cf] probably derived from desiccation and brecciation of mudcracks. Scale bar is 15 cm. (C) Infilled cast of a hopper crystal. Coin is 2.1 cm. (D) Vertically stacked tidal-flat sequences [bracket] consisting of subtidal massive sandstones [lithofacies Smc], intertidal sandstones [lithofacies Smf], and mud drapes [lithofacies Fm], and overlying supratidal carbonates [lithofacies Cm]. The carbonates at this location have been dolomitized. Scale bar is 1 m. (E) Nodular carbonate [lithofacies Cn] indicated by arrows, overlain by and with infilling arenaceous carbonate mudstone [lithofacies Cm]. These stratiform breccias are interpreted as surficial paleokarst collapse features overlying dissolved stratiform evaporites. Scale bar is 15 cm. (F) Synsedimentary deformation in carbonates [lithofacies Cn] interpreted as relict structures from enterolithic folding in replaced evaporites. Scale bar is 15 cm. (G) Interpreted teepee structure in supratidal carbonate [lithofacies Cn]. Scale bar is 15 cm.
syndepositional folds. The combination of structures is consistent a carbonate peritidal environment representing upper intertidal and supratidal zones (Pratt, 2010). The presence of evaporites, sediment reddening, and incorporated eolian quartz grains suggests that arid or semiarid paleoclimate conditions affected the uppermost intertidal and supratidal zones.

**Facies Association 7 (FA7): Estuarine-Mouth Tidal-Sandbar Deposits**

*Description.* FA7 is composed of sandstone packages 1–2 m thick with low-angle accretional surfaces (Fig. 10A). These are internally organized into basal granule lag deposits, overlain by stacked sets of trough cross-beded sandstone, and capped by wave- or current-rippled sandstone (Fig. 10B). Individual cross-bed sets are typically <10 cm thick and are separated by internal erosion surfaces. Cosets consist of packages of three to five cross-bed sets separated by partial to complete mudstone laminae (Fig. 10C). These delineate tidal bundles separated by reactivation surfaces (e.g., Greb and Archer, 1995). There are rare examples of herringbone cross-bedding (Fig. 10D) and wave ripples (Fig. 10E) at the tops of cosets. The tops of sandstone bodies are commonly intensely bioturbated omission surfaces dominated by *Planolites* (Fig. 10F) or *Lockeia* (Fig. 10G).

*Interpretation.* The larger sandstone bodies are interpreted as estuarine-mouth tidal-sandbar deposits, showing evidence for tidal current reversals and reworking in the form of reactivation surfaces, internal erosion surfaces, herringbone cross-bedding, wave ripples, and partial to complete mud drapes (Dalrymple, 2010). These are similar to successions observed in modern estuary-mouth settings (Clifton and Phillips, 1980). In some cases, these bar deposits are interbedded with estuarine central basin deposits. In other cases, bar deposits transition upward to siliciclastic tidal-flat deposits (Fig. 4). Intensely bioturbated horizons that are interpreted as firmgrounds that developed in bar-top settings and were re-exhumed by erosional loss of estuarine basin muds that originally draped the bedform.

Alternative explanations include a barrier sequence, or inlet channels with related flood-tide or ebb-tide deltas. The barrier interpretation is rejected in this case because these deposits do not resemble back-barrier spillover lobes, lacking significant upper-flow-regime plane-bed structures or foresets that would be in accord with transgressive barrier retreat into estuarine central basin settings. There is an absence of characteristic beach, nearshore bar, rip current, or shoreface deposits (e.g., Castle, 2000). The lateral migration of inlet channels could possibly produce similar channel bedforms and successions (e.g., Moslow and Tye, 1985), but again there is no evidence supporting the presence of estuarine mouth-barrier deposits. Ebb-tide deltas can be excluded because there is an absence of interbedded offshore deposits and features. Flood-tide deltas that prograded into the central estuarine basin remains a possible alternative explanation, but is judged less likely due to a lack of evidence for significant storm-wave reworking (e.g., Boyd, 2010).

**Provenance, Paleocurrents, and Source-Area Evolution**

**Sediment Composition**

*Description.* Sandstone in the Ignacio Formation is mostly quartz arenite and quartz wacke (Table S1 [footnote 1]), with typically 75% ± 15% monocrystalline quartz and 20% ± 13% polycrystalline quartz (Fig. 11A). Some locations have extremely well rounded, “billiard ball” quartz (Fig. 11B), attributed to eolian input by McBride (2016a). There are, however, individual samples with up to 32% feldspar (average 4% ± 8%), including both fresh and highly altered individual grains of microcline (Fig. 11C) and plagioclase. Lithics typically comprise <5%. Rare lithic clasts include granite (Fig. 11D), quartzite (Fig. 11E), chert (Fig. 11F), and mudrocks (Fig. 11G); all of these except chert had local sources. Carbonate deposits are arenaceous or argillaceous limestone or dolostone (Fig. 11H), and the incorporated well-rounded quartz sand or angular-subangular quartz silt is consistent with eolian input.

*Interpretation.* Sandstone point counts plot in the quartzose recycled petrofacies field of monocrystalline quartz–total feldspar–total lithics (QmFlt) diagrams (Fig. 12). This is interpreted as reflecting the composition of adjacent source rocks: Proterozoic granite (granitic rock fragments and microcline), quartzite from the Proterozoic Uncompahgre Formation, and plagioclase from the Proterozoic Twilight Gneiss. Rare sandstone and mudrock lithics may have sourced from the East Lime Creek Conglomerate. The source of the chert lithics is not presently known, but one possible source would be the Ptarmigan Chert Member of the Lower Ordovician Manitou Formation.

**Paleocurrents**

*Description.* Paleocurrent analysis is based on 124 cross-bed measurements and 24 crestline orientations from wave ripples (Fig. 13). The cross-bedding data show a statistically significant (p < 0.05) unimodal pattern with relatively high statistical dispersion. The vector mean is to the northwest (azimuth 320°). Crestlines of wave ripples are oriented ESE-WNW (azimuth 105°–285°), which is interpreted to show NNE-SSW (azimuth 015°–195°) paleoflow directions. However, when it was possible to measure paleocurrent data from both cross-bedding and ripple marks at the same locality, the paleoflow directions tended to be virtually identical (Fig. 13). Across the study area, paleoflow direction appears to shift from north directed in the southern part to more west and northwest directed elsewhere (Fig. 13).

*Interpretation.* The paleocurrent patterns are consistent with an estuarine depositional system. Bedload transport in modern estuaries is typically ebb-tide oriented throughout the entire estuary, and flow reversals are usually insufficient to radically affect the dominant orientation of larger estuarine bedforms (Home and Patton, 1989). In this case, paleoflow in the fluvial and proximal estuarine part of the study area is west or northwest directed with a higher range of statistical dispersion. This is interpreted as flow in sinuous streams (higher...
Figure 10. Field photographs of estuarine-mouth tidal sandbars (facies association FA7; see Table 3). (A) Macroform (tidal sandwave) overlying central estuarine basin mudstones (lithofacies Fm; see Table 2). Person for scale. (B) Multistory packages of trough cross-bedded sandstones (lithofacies St) and wave-rippled sandstones (lithofacies Srw) with internal erosion surfaces and partial mud drapes (lithofacies Fm). Hammer is 28 cm. (C) Reactivation surface in cosets of trough cross-bedded sandstones. Scale bar is 5 cm. (D) Herringbone cross-bedding (lithofacies Sph) and overlying wave ripples (lithofacies Srw). Scale bar 15 cm. (E) Wave ripple marks (lithofacies Srw) on an exposed reactivation surface within a tidal sand-bar sequence. Pen is 14 cm. (F) Firmground dominated by *Planolites* (*P*). Scale bar is 7 cm. (G) Firmground dominated by *Lockeia* (*Lo*) and *Planolites* (*P*). Scale bar is 7 cm.
Figure 11. Photomicrographs from the Ignacio Formation, showing that the unit is dominantly subrounded to subangular monocrystalline quartz (A) but also includes “billiard ball” well-rounded quartz (B) (scale bar is 250 μm in both images). Ignacio Formation from locations near granitic source areas can have significant amounts of microcline (C; scale bar is 250 μm) and granitic rock fragments (D; scale bar is 500 μm). Quartzite (metamorphic rock fragments) (E) is probably sourced from the Uncompahgre Formation (scale bar is 500 μm). Chert (F) is a minor constituent, and the source area is not known (scale bar is 500 μm). Also shown are a sedimentary rock fragment (mudstone) with ferruginous cement from intertidal zone deposits (G; scale bar is 500 μm) and an arenaceous carbonate mudstone from supratidal zone deposits (H; scale bar is 250 μm).
Source-Area Evolution

Description. At Sultan Creek (location 17 in Fig. 3), a stratigraphic succession of five samples was evaluated for detrital zircon U-Pb geochronology. The lowest sample, a sandstone in the East Lime Creek Conglomerate, has an age spectrum peak of ca. 1.7 Ga (Fig. 14). In the overlying Ignacio Formation, the four samples show the continuity of the ca. 1.7 Ga age peak, but also show a stratigraphically progressive development of a ca. 1.4 Ga age peak, producing a bimodal pattern near the top of the unit (Table S2 [footnote 1]). Similar detrital zircon U-Pb age results were obtained by McBride (2016a), although his samples were not collected from a continuous stratigraphic sequence.

Interpretation. The Proterozoic basement consists of the Twilight Gneiss, the Uncompahgre Formation, and a series of granitic intrusions. The protolith of the Twilight Gneiss was bimodal dacitic-hypabyssal volcanic rocks and associated pelitic rocks that were deformed and metamorphosed to hornblende-plagioclase gneiss and amphibolite (Barker, 1969) as part of a regional collision event (Yavapai orogeny) between the Mojave and Yavapai terranes (Gonzales and Van Schmus, 2007; Amato et al., 2008; Jones et al., 2009). Concurrent with terrane accretion, the Twilight Gneiss was intruded by 1.72–1.68 Ga trondhjemite plutons (Barker, 1969; Hutchinson, 1976; Baars et al., 1987; Gonzales and Van Schmus, 2007; Jones et al., 2009).

The protolith of the 2.5-km-thick Uncompahgre Formation was quartz arenite with minor shale and conglomerate (Harris and Eriksson, 1990), with a depositional age ca. 1.7 Ga (Jones et al., 2009). The Uncompahgre Formation underwent multiple phases of deformation linked to the collision of the Yavapai and Mazatzal terranes (Mazatzal orogeny) starting ca. 1.66 Ga (Shaw and Karlstrom, 1999; Amato et al., 2008; Jones et al., 2009), then was buried to crustal depths of 10–15 km, metamorphosed to greenschist facies, and intruded by the 1.44 Ga Eolus Granite (Bickford et al., 1969; Hutchinson, 1976; Gonzales and Van Schmus, 2007; Jones et al., 2009).

The contact relationship between the Twilight Gneiss and Uncompahgre Formation has been controversial, with explanations ranging between: a depositional contact (Barker, 1969; Gibson and Harris, 1992); a high-angle fault contact (Baars et al., 1987); the Uncompahgre Formation being thrust southward over the Twilight Gneiss (Tewksbury, 1985); and a model where the Uncompahgre Formation was initially deposited above the Twilight Gneiss, then “decoupled” and thrust northward (Harris et al., 1987). However, field data show that the Twilight Gneiss was thrust northward over the Uncompahgre Formation, based on the following evidence: (1) klippen of gneiss sitting on top of the Uncompahgre Formation; (2) sheath folds in the gneiss where it ramps over the quartzite; (3) fault-zone cataclastic metamorphism on the top of the Uncompahgre Formation (forming a carapace of shattered purple quartzite encased in white vein quartz at the top of the lower plate); and (4) retrograde metamorphism from gneiss to schist at the base of the Twilight Gneiss, representing deformation throughout the lower part of the upper plate (Evans and Holm-Denoma, 2018).

Ignacio Formation detrital zircon U-Pb geochronology data support this new thrust-fault interpretation. The initial ca. 1.7 Ga age spectrum peak represents erosion of the Twilight Gneiss from the allochthonous upper plate, while the ca. 1.4 Ga age spectrum peak represents erosion of the Uncompahgre Formation from the parautochthonous lower plate. Thus, the stratigraphic trend, showing the initial appearance of the ca. 1.7 Ga age spectrum peak and then the subsequent development of the ca. 1.4 Ga age spectrum peak, can be interpreted to show the unroofing history of this Proterozoic collisional zone (Fig. 14).

DISCUSSION

Tide-Dominated Estuarine Depositional System

The distinguishing characteristic of the Ignacio Formation is the prevalence of tidal sedimentary structures including tidalites with double mud drapes; flaser, wavy, and lenticular bedding; cross-bedded sandstone with mudstone drapes and reaction surfaces; sandstone with herringbone cross-bedding;
mudstone intraclasts derived from mudcracked or brecciated muds; and carbonate-evaporite tidal-flat deposits with crystal casts, teepee structures, enterolithic folding, surficial karst, breccias from stratiform evaporite dissolution, and nodules. Several facies associations demonstrate the importance of tidal processes, including tidally influenced fluvial environments, siliciclastic and carbonate tidal flats and supratidal flats, and estuarine-mouth tidal sandbars that demonstrate remolding by tidal current reversals. Accordingly, the Ignacio Formation estuarine depositional systems is considered tide dominated in the classification scheme of Ainsworth et al. (2011).

Diagnostic estuarine subenvironments are bayhead deltas, estuarine central basins, and estuarine channels. In the Ignacio Formation, bayhead deltas are marked by IHS deposits, distributary mouth-bar successions, and wave-modified turbidites. Estuarine central basin deposits are sandstone rich, which, coupled with the unimodal west-directed paleocurrent pattern, suggests significant fluvial bedload transport east to west through the full length of the estuary, similar to the modern Connecticut River estuary (northeastern USA) (Horne and Patton, 1989). The pervasive red colors in the fluvial and bayhead delta components of this depositional system may be indicative of fluvial transport of large amounts of red sand and mud into the proximal reaches of an estuarine depositional system, again showing the importance of the fluvial component of the depositional system. In summary, the estuarine systems in the Ignacio Formation are considered fluvial influenced as a modifier of the tide-dominated classification.

Figure 13. Map showing the locations of paleocurrent data. Single arrows indicate the azimuth of unidirectional measurements (cross-bedding), and double arrows indicate the azimuths for bidirectional measurements (wave ripples). Paleocurrent rose diagrams summarize the cross-bedding and wave-ripple data. Abbreviations: CBP—Coal Bank Pass; Ck.—Creek; n—number of measurements; VM—vector mean; csd—circular standard deviation (Krause and Geijer, 1987); p—Rayleigh test of significance (Curray, 1956). For bidirectional data, the mode identifies the probable mean paleocurrent direction.
Figure 14. Stratigraphic section at location 17 (see Table 1 and Figs. 1 and 3 for location and symbol explanation) showing the stratigraphic position of a series of detrital zircon U-Pb age spectra determinations (sample numbers in upper right). The vertical axis is the number of individual grains (blue histogram) and relative probability (red curve), while the horizontal axis is preferred age in millions of years (Ma). The data show a stratigraphic trend from a unimodal 1.7 Ga age peak at the base of the Ignacio Formation, to bimodal 1.4 Ga and 1.7 Ga age peaks. These data are interpreted to show the unroofing history of a Proterozoic thrust fault system, specifically that the allochthonous upper plate (1.7 Ga Twilight Gneiss) was eroded prior to exposure of the parautochthonous lower plate (1.4 Ga Uncompahgre Formation). See text for details. Cgl.—Conglomerate.
The importance of wave processes is suggested by wave ripples, wave modification of delta-front turbidites, locally developed micro-hummocky stratification, and evidence for resuspension of estuarine mud to produce lag surfaces, omission surfaces, and firmgrounds. However, evidence of strong wave action is lacking (e.g., Vakarelov et al., 2012). For example, there are only a few occurrences of tempestites, and there is no evidence for estuary-mouth constructional sand bodies such as spits or barriers (Dalrymple, 2010). Accordingly, the term wave affected becomes the final modifier of this estuarine depositional system classification.

As a tide-dominated, fluvial-influenced, wave-affected (Tfw) estuarine depositional system, the depositional environment of the Ignacio Formation compares to similar modern estuaries undergoing transgression. There is an east-west trend of fluvial deposits to increasingly tide- and wave-produced features and deposits, similar to the proximal-to-distal relationships seen in modern estuaries (e.g., Dalrymple et al., 1990). In addition, there is a stratigraphic trend from fluvial deposits to estuarine deposits to nearshore marine deposits (Fig. 15). This trend is interpreted as an overall transgressive succession. This interpretation is supported by the presence of numerous tidal-fluvial ravine-ment surfaces within the estuarine central basin deposits (e.g., Boyd, 2010).

### Incised Valley Fills

Incised valley systems (IVSs) consist of paleovalleys created during prior base-level fall, then backfilled by fluvial, estuarine, and coastal depositional systems during subsequent base-level rise (Allen and Posamentier, 1993). Evidence for IVSs in the geological record requires (1) erosion of underlying geologic units (Zaitlin et al., 1994); (2) regional significance of the basal erosion.

![Diagram showing the thickness and lateral distribution of deposits of the depositional environments in the Ignacio Formation estuarine depositional system.](https://pubs.geoscienceworld.org/gsa/geosphere/article-pdf/15/5/1479/4831251/1479.pdf)
Incised valley-fill successions containing estuarine and bayhead-delta deposits have been recognized from both modern and ancient settings as markers of lowstand, transgressive, and early highstand systems tracts (Aschoff et al., 2018; Simms et al., 2018). Three principles are used in this analysis: (1) the IVS basal unconformity is interpreted as a sequence boundary (SB) representing relative sea-level fall, and exposure and erosion of continental and nearshore marine deposits (Miller et al., 2013); (2) IVS estuarine depositional systems commonly include numerous of fluvial-tidal ravinement surfaces (Frey and Howard, 1986); and (3) the transition from estuarine depositional systems to embayment marine facies is a maximum flooding surface (MFS), typically located at the top of the incised paleovalley infill (Miller et al., 2013).

As discussed, there is significant evidence for an erosional unconformity at the base of the Ignacio Formation, where fluvial deposits overlie marine deposits of the East Lime Creek Conglomerate. This basal unconformity matches criteria used by other studies (e.g., Miller et al., 2013) and is interpreted as a SB (Fig. 16). Above the SB, the Ignacio Formation is restricted to incised paleovalleys: at different locations, each of the different facies components of the Ignacio Formation estuarine depositional system onlaps the paleovalley walls. The units are unrelated because trends in facies distributions or paleocurrent directions do not match. Finally, basin-fill onlap is recorded in lateral trends in paleovalleys (Fig. 15).

Thickness variations within the Ignacio Formation delineate at least three paleovalleys (labeled 1–3 in Fig. 15) in the study area. The dimensions of the paleovalleys are difficult to estimate because of inadequate geographic distribution of outcrops, resulting in unknowns such as paleovalley shape, the paleovalley position of any measured section, and whether or not paleovalleys amalgamate basinward. The paleovalleys were ≥42 m deep and probably up to 10–30 km in width. The infill of each paleovalley shows a consistent stratigraphic succession of fluvial, tidally influenced fluvial, bayhead delta, estuarine central basin, and estuary-mouth tidal-sandbar deposits.

### Sequence Stratigraphy

Sequence stratigraphy is the study of the spatial and temporal relationships of sedimentary rocks, particularly sediments deposited in marine basins. It is a framework for understanding the geological history of an area over a long period of time. In the context of the Ignacio Formation, the sequence stratigraphy helps to interpret the depositional history of the area, including the timing and nature of sea-level changes and the deposition of sediments in different environments.

In the case of the Ignacio Formation, the sequence stratigraphy includes the identification of unconformities and the correlation of depositional units across the study area. The sequence stratigraphic framework includes the identification of sequence boundaries, transgressive systems tracts, and highstand systems tracts. These units are recognized based on changes in facies and sedimentology, as well as changes in sea-level history.

**Figure 16. Interpretation of the composite sequence stratigraphy of the Ignacio Formation.** At different locations, different components of the Ignacio Formation estuarine depositional system overlie and onlap onto Proterozoic crystalline basement and/or the East Lime Creek Conglomerate. The evidence for an unconformity between the Ignacio Formation and East Lime Creek Conglomerate is discussed elsewhere (Evans and Holm-Denoma, 2018). The upper member of the Elbert Formation also overlies Proterozoic crystalline basement and/or the East Lime Creek Conglomerate at different locations. This basal unconformity is interpreted as a sequence boundary (SB). Fluvial, tidally influenced fluvial, and bayhead delta facies are interpreted as a lowstand systems tract (LST). The upper boundary of the LST both is a fluvial-tidal ravinement surfaces marked by a pebble lag surface, and represents a flooding surface that puts estuarine facies above fluvial facies; accordingly, this surface is interpreted as a transgressive surface of erosion (TSE). The upper part of the Ignacio Formation is estuarine central basin, estuarine channel, tidal flat, and estuary-mouth tidal sandbar deposits. This is interpreted as a transgressive systems tract (TST). The maximum flooding surface (MFS) is placed at the transition from a fluvial-estuarine system confined by paleotopography, to an embayment depositional system that is unconfined. This corresponds to the Ignacio Formation–Elbert Formation (upper member) contact. The upper member of the Elbert Formation and the Ouray Limestone (Dyer Formation) is interpreted as a highstand systems tract (HST). There is a significant unconformity separating the Ouray Limestone from the Leadville Limestone, interpreted as another sequence boundary. Lithology symbols are as shown in Figure 3.
to unconformably overlie the Proterozoic crystalline basement and/or the East Lime Creek Conglomerate.

The lower part of the Ignacio Formation consists of aggradational to retogradational stacked fluvial, tidally influenced fluvial, and bayhead delta facies. The upper contact of the bayhead delta facies is typically marked by a fluvial-tidal ravinement surface, representing delta lobe abandonment and wave reworking. This first flooding surface corresponds to the onset of marine conditions in the paleovalleys, transitioning to estuarine conditions (e.g., Rossi et al., 2017). Because this surface corresponds to both reworking and flooding, it is interpreted as the transgressive surface of erosion (TSE) marking the top of the lowstand systems tract (LST).

The upper part of the Ignacio Formation consists of retogradational stacked estuarine channel fills, central basin deposits, and estuary-mouth tidal sand-bar deposits. There are numerous fluvial-tidal ravinement surfaces (F/TRS) and firmgrounds representing the effect of river floods or enhanced tidal currents on estuarine deposition. The upper boundary is placed at the conformable contact between the Ignacio Formation and overlying upper member of the Elbert Formation. This lithological transition from marginal marine, mixed siliciclastic-carbonate deposition confined to paleovalleys, to marine, carbonate-dominated deposition unconstrained by paleotopography is interpreted as the MFS, and probably represents marine onlap onto interfluve areas of the underlying paleotopography (e.g., Rossi et al., 2017). Thus, the upper part of the Ignacio Formation is a transgressive systems tract (TST).

The upper member of the Elbert Formation is poorly exposed, but likely forms part of an extensive marine embayment in conjunction with offshore carbonate of the overlying Ouray Limestone. Other studies have found that continued transgression transitions a funnel-shaped fluvial-estuarine system to a coastal embayment, with proximal marine mud and a distal carbonate platform (Fischbein et al., 2009; Aschoff et al., 2018). Thus, the upper member of the Elbert Formation and Ouray Limestone is interpreted as a highstand systems tract (HST). There is a significant erosional unconformity separating the Ouray Formation from the overlying Mississippian Leadville Limestone (Klink et al., 2013), which is probably another SB. At sites throughout western Colorado and Utah, Myrow et al. (2013) found palaeokarst in the upper Dyer Formation (equivalent to the Ouray Limestone) and isotopic evidence linking this unconformity to the latest Famennian Hangenberg event, interpreted as an eustatic sea-level fall event, and subsequent development of lowstand and transgressive systems tracts.

## SUMMARY AND CONCLUSIONS

The lower Paleozoic (Cambrian–Devonian) history of the San Juan Mountains in southwestern Colorado has been poorly understood for the past century. Recent revisions of the stratigraphic framework include recognition that the basal conglomerate formerly of the Ignacio Formation is a separate unit (Evans and Holm-Denoma, 2018); recognition of an unconformity between this unit and the Ignacio Formation; and reassignment of strata traditionally called the McCracken Sandstone Member of the Elbert Formation to the Ignacio Formation. In addition, new fossil evidence clarifies that the Ignacio Formation is Upper Devonian, and detrital zircon U-Pb geochronology data are consistent with that age assignment.

Facies analysis shows that the Ignacio Formation was a tide-dominated, fluvial-influenced, wave-affected, transgressive estuarine depositional system that backfilled at least three incised paleovalleys in western Colorado. The basal unconformity represents a sequence boundary. The fluvial-tidal ravinement surface directly above the bayhead delta facies represents a transgressive surface of erosion, and thus the lower fluvial part of the Ignacio Formation is an Upper Devonian lowstand systems tract. The overlying estuarine central basin and estuary channel deposits, tidal-flat deposits, and estuarine tidal sand-bar deposits are capped by a maximum flooding surface representing a change from marginal marine deposits confined by paleovalleys to marine embayment deposits (upper member of the Elbert Formation and Ouray Limestone) not confined to paleovalleys, thus the upper part of the Ignacio Formation is an Upper Devonian transgressive systems tract, and the overlying units are part of a highstand systems tract.

Finally, local sediment sources for the Ignacio Formation were controlled by the erosion and unroofing history of a Proterozoic thrust-sheet complex. Additionally, the presence of evaporites, eolian sediment input, and source-area terrestrial reddening suggest that the surrounding region was affected by arid to semiarid paleoclimate.

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