Assessment of backwater controls on the architecture of distributary channel fills in a tide-influenced coastal-plain succession: Campanian Neslen Formation, USA

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ABSTRACT

The backwater zone of a river is its distal reach downstream of the point at which the streambed elevation reaches the sea level. Backwater hydraulics is believed to exert an important control on fluvio-deltaic morphodynamics, but the expressions with which this may be recorded in the preserved stratigraphic record are not well understood. The seaward reaches of modern rivers can undergo flow acceleration and become erosional at high discharges due to drawdown of the in-channel water surface near the river mouth, in relation to the fixed water surface at the shoreline. As coastal-plain distributary channels approach the shoreline they commonly tend to be subject to a reduction in lateral mobility, which could be related to diminished sediment flux at low flow.

Current understanding of channel morphodynamics associated with backwater effects, as based on observations from numerical models and modern sedimentary systems, is here used to make predictions concerning the architecture of coastal distributary channel fills in the rock record. On the basis of existing knowledge, distributary channel fills are predicted to be typically characterized by low width-to-thickness aspect ratios, by a clustering of scour surfaces toward their base, by an aggradational infill style, by a facies organization that bears evidence of drawdown-influenced scour filling, possibly resulting in the overprint of tidal signals toward their base, and by co-genetic sand-prone overbank units of limited occurrence, thickness and sand content.

To test these predictions, fieldwork was carried out to examine sedimentological characters of channel bodies from an interval of the Campanian Neslen Formation (eastern Utah, USA), which comprises a succession of sandstone, carbonaceous mudstone, and coal, deposited in a coastal-plain setting, and in which significant evidence of tidal influence is preserved.

Three types of channel bodies are recognized in the studied interval, in terms of lithology and formative-channel morphodynamics: sand-prone laterally accreting channel elements, heterolithic laterally accreting channel elements and sand-prone aggradational ribbon channel elements. This study concentrates on the ribbon channel bodies since they possess a geometry compatible with laterally stable distributaries developed in the zone of drawdown. Sedimentological and architectural characteristics of these bodies are analyzed and compared with the proposed model of distributary channel-fill architecture.

Although conclusive evidence of the influence of backwater processes in controlling the facies architecture of distributary channel fills is not reached, the studied bodies display an ensemble of internal architecture, lithological organization, nature of bounding surfaces and relationships with other units that conforms to the proposed model to a certain extent. The analyzed ribbon sandbodies are all characterized by erosional cut-banks, very limited proportions of mudstone deposits, a lack of genetically related barform units, clustering of scour fills at their base, and a lack of relationships with co-genetic river-fed overbank sandstones.

This work provides a guide to future research, which is required to better understand the role of backwater processes in controlling the architecture of distributary channel bodies, their down-dip variations, and how these are expressed in the stratigraphic evolution of prograding coastal plains.

Keywords: distributary; channel; delta; backwater; morphodynamics.

INTRODUCTION

Background

The backwater zone of a river is defined as the distal reach where the streambed drops below sea level resulting in river-flow deceleration on approach to the static water body into which it discharges (cf. Chatanantavet et al. 2012; and references therein). Significant recognition is now given to the role of backwater hydraulics as a control on fluvio-deltaic morphodynamics (Chatanantavet et al. 2012; Lamb et al. 2012; Nittrouer et al. 2012; Chatanantavet & Lamb 2014; Ganti et al. 2014), and this has raised awareness of its potential importance as a factor controlling sedimentary architecture in the preserved stratigraphy of corresponding preserved successions (Lamb et al. 2012; Blum et al. 2013).

Fundamentally, backwater processes are considered to exert a control on the location of avulsion nodes that dictate the size of deltas and the distributary nature of their channel patterns. Therefore, these processes determine – at least in part – the fundamental length-scale of deltaic systems (Chatanantavet et al. 2012; Ganti et al. 2014). In addition, backwater effects are considered to act as a filter on source-to-sink sediment flux, by muting bedload flux through enhanced storage in the upper backwater zone during low flows, and by increasing bedload flux through re-mobilization of this stored sediment during high flows (Lamb et al. 2012; Nittrouer et al. 2012). This transition from depositional to erosional behavior is associated with a process called water-surface drawdown, a mechanism whereby the distal reach of a river becomes erosional during episodes of high discharge due to drawdown of the fluvial water surface near river mouth, which results in flow acceleration toward the shoreline. This process is enabled because the river plume spreads laterally beyond the shoreline, thereby rendering the plume surface relatively fixed in elevation (Lane 1957; Lamb et al. 2012). Evidence for the effectiveness of water-surface drawdown as a geomorphic driver originates from observations from modern fluvio-deltaic systems and from numerical models, which also indicate that the length over which drawdown propagates upstream is a function of flood discharge, and can be larger than half the total backwater length for major floods (Lamb et al. 2012). The combination of sediment starvation at backwater and scouring at drawdown provides an explanation for the occurrence of distal backwater zones characterized by laterally stable, incisional distributary channels (cf. Kolb 1963; Hudson & Kesel 2000; Gouw & Berendsen 2007; Jerolmack & Mohrig 2007; Jerolmack 2009; Nittrouer et al. 2012).

Although backwater hydraulic conditions have been suggested to have an influence on channel behavior and morphology across present-day delta plains, hitherto there has been only modest consideration of how these processes may be recorded in the stratigraphic record (Fig. 1a). Specifically, consideration of how backwater hydraulics might have influenced the architecture of fluvio-deltaic sedimentary successions has mostly been concerned with assessment of down-gradient variations in channel sandstone geometries. Blum et al. (2013) noted the expected downstream evolution in the deposits of the modern Mississippi River, in which interpreted channel sandbodies are observed to widen down-system until reaching the uppermost end of the backwater zone, after which they narrow (Fig. 1a). However, more comprehensive evidence is currently lacking to support the importance of backwater hydraulics as the dominant control on channel mobility. Petter (2010) interpreted a down-system increase in inferred paleo-flow depth for channel bodies in the Cretaceous Lower Castlegate Formation in Utah (USA) and related this to forcing by backwater conditions (Fig. 1a); his observations are in agreement with the expected behavior of channels developed in the zone of water-surface drawdown (cf. Lamb et al. 2012).

Based on observations from modern rivers (e.g. Choi et al. 2004; Van den Berg et al. 2007; Martinius & Van den Berg 2011; Johnson & Dashtgard 2014; La Croix & Dashtgard 2014; 2015) and ancient successions (e.g. Shanley et al. 1992; Martinius & Gowland 2011; Martinius & Van den Berg 2011; Martinius 2012), a number of authors have studied the lithological characteristics of channel deposits in the fluvial-tidal transition zone – which overlaps with the backwater zone – to establish facies criteria for the interpretation of the rock record. Although these works do not particularly focus on the potential influence of backwater and drawdown hydrodynamics, they provide a basis for the discernment of depositional patterns that may arise in response to backwater controls on tidal-fluvial interaction.

Sambrook Smith et al. (2010) argue that the sedimentological imprint left by extreme floods in the facies organization of alluvial channel fills is not expected to differ significantly from the signature left by minor floods, as the channels respond to larger floods by over-topping their banks rather than deepening. However, this type of response may not be the rule for distributary channels subject to water-surface drawdown during flood events. This might result in the accumulation of a particular arrangement of lithofacies with a recognizable style of internal organization within preserved channel fills that developed under such conditions. Similarly, it is also of significant importance to determine if and how backwater processes may interact with tidal processes in determining the facies organization of distributary channels in the fluvial-tidal transition zone of tidally influenced deltaic settings. Changes in lithofacies characters through the fluvial-to-marine transition zone are believed to record a seaward decrease in the intensity of river flow and a seaward increase in the intensity of tidal currents. This concept is embedded in facies models for distributary channels (e.g. Dalrymple & Choi 2007), which do not account for potential interference exerted by the process of water-surface drawdown, whose magnitude also decreases upstream. These facies models give no consideration, for instance, to the possibility that the process of water-surface drawdown may scour the distributaries deep enough to permit the deposition of flood-related deposits, which may leave no evidence of tidal influence due to the overriding fluvial input that prevails during episodes of high-discharge flood events. In this scenario, the flood peak would presumably be expressed as an erosive scour surface, whereas the receding limb of the hydrograph would be recorded in deposits that fill some portion of the scour. Such deposits would likely have high preservation potential due to their accumulation in the lowermost parts of deep channel scours.

On the basis of observations from modern depositional systems and results from numerical models (Chatanantavet et al. 2012; Lamb et al. 2012; Blum et al. 2013; Chatanantavet & Lamb 2014), it is possible to hypothesize the influence that backwater hydraulics and the water-surface drawdown process potentially exert on the sedimentary architecture of distributary-channel fills in lower delta-plain settings, in terms of geometries, internal organization and relationships with neighboring sedimentary units. These features are expected to relate to the particular behavior of the formative channels, both during major high-discharge events and in the longer term (e.g. the planform evolution of these bodies). For the sake of conceptualization, two end-member behaviors can be envisaged by contrasting channels in alluvial plains, upstream of backwater influences, with channels developed at maximum drawdown (Fig. 1b). Channels in alluvial plains are expected to have an 'ordinary' response to large floods, whereby the streambed undergoes some degree of incision and the channel water surface rises, overtopping the levees and resulting in flooding of overbank regions (cf. Sambrook Smith et al. 2010). In contrast, at its mouth, a distributary channel is ideally expected to respond to large river floods solely by deeply scouring its bed (cf. Lamb et al. 2012). A distributary channel developed in the zone of drawdown may have a behavior that approximates the latter end-member to some degree, so that the frequency and magnitude with which river-driven floods act upon adjacent coastal-plain areas may be diminished with respect to up-dip alluvial plains. In this domain, high-tide floods overtop channel margins and cause overland flow and associated overbank accretion (cf. Eisma 1997): levee-sandstone development is likely inhibited particularly where tidal range is large relative to fluctuations in water-surface height induced by river floods, because high-tide flows that overtop channel banks tend to be weak (Allen & Chambers 1998). More generally, the reduced water-surface elevation connected with drawdown during floods, potentially in combination with a depression of the in-channel sediment-concentration profile associated with channel deepening, could affect the export of sand to the proximal overbank on a delta plain through a control on advection and turbulent diffusion processes (Pizzuto 1987, and references therein; Adams et al. 2004). In view of this, it is possible that drawdown hydrodynamics contribute to the seaward decrease in the relief of channel banks seen in modern systems (cf. Hill et al. 2001; Fielding et al. 2005; Funabiki et al. 2012).

These theoretical predictions of the sedimentological characteristics of lower-delta-plain distributary channels that consider the possible influence played by water-surface drawdown are here assessed against purposely acquired field data. The sedimentological characteristics that are thought to potentially represent indicative criteria of the influence of backwater processes on distributary channel fills are summarized by the following points.

- External and internal geometries of the distributary-channel fills and the nature of their banks are expected to reflect the low-sinuosity channel pattern, which itself relates to the reduced channel mobility, and the effect of high-flow punctuation. Therefore channel fills are predicted (i) to exhibit ‘ribbon’-like external geometries (*sensu* Friend, 1983) characterized by low width-to-thickness aspect ratios (indicatively less than 15; Friend 1983), and (ii) to be internally characterized by a multi-story aggradational style of infill that comprises a series of erosively based bedsets with a broadly vertical or concentric (*sensu* Hopkins, 1985) style of stacking. Given the expected geomorphic stability of distributary channels in the lower backwater zone, the clustering of high-relief scour surfaces toward the base of narrow channel fills is deemed more likely than the formation of laterally extensive erosional surfaces that can be misinterpreted as sequence boundaries (cf. Lamb et al. 2012).
- The facies organization of the channel fills may be variably affected, but overall there is reason to consider the likelihood of systematic occurrence of lowermost channel storys that represent a record of the peak (the erosive base) and possibly of the receding limb (the overlying scour infill) of major flood events. Notably, this predicted

character would be expected to be seen also in tidally influenced depositional systems, which may be characterized by distal distributary channel fills with lowermost portions that carry no evidence of tidal influence, particularly if bank accretion does not take place. Such a depositional style would occur where the tidal limit had been temporarily displaced seaward by a dominant fluvial signal during the early stage of the falling hydrograph limb. Fluvially dominated deposits may constitute a large part of the infill of such channels because of the high preservation potential of the drawdown scour, but only if the streambed elevation responds in phase with discharge variations (cf. Chatanantavet & Lamb 2014).

- Horizontal spatial relationships with other sedimentary bodies are expected to reflect the incisional character of channels developed in zones of drawdown hydrodynamics and possibly diminished overbank flow. The relative rate of accumulation of organic material compared to fine-grained clastic detritus in regions adjacent to distributary channels may be enhanced, and the likelihood of development of co-genetic sand-rich levees or crevasse splays may be significantly diminished relative to upstream alluvial settings. This could be reflected in a downstream decrease in the frequency, thickness and sand content of proximal overbank sandstones in deposits preserving a record of the zone of water-surface drawdown.

Aims and Objectives

The principal aim of this work is to assess the potential validity of the proposed ideas against field data from a suitable tidally influenced coastal-plain depositional system.

Specific research objectives of this work include:

- the description of channel bodies from a suitable field case study, and interpretation of the processes they record;
- the assessment of the degree to which the proposed criteria for the recognition of the influence of backwater processes on low-sinuosity distributaries in tidally influenced

settings are seen to co-occur in the studied succession; this assessment is based on the evaluation of the architecture, facies organization and relationships with overbank deposits displayed by distributary channel fills that have internal and external geometries compatible – though not in themselves diagnostic – with how backwater hydraulics are thought to control channel morphodynamics;

- a discussion of the likelihood of preservation of backwater signals in the studied succession.

FIELD CASE STUDY

Field-data collection has been undertaken to analyze a stratigraphic interval of the Campanian Neslen Formation, which forms part of the Upper Cretaceous Mesaverde Group and which crops out along the Book Cliffs in eastern Utah, USA (Fig. 2a; 3). The studied interval is part of a thick succession that accumulated as an eastward-prograding clastic wedge in a retroarc foreland basin on the western margin of the Western Interior Seaway (Lawton 1986; Miall et al. 2008, and references therein). The Mesaverde Group succession has been extensively studied (e.g. Lawton 1986; Olsen et al. 1995; Van Wagoner 1995; Yoshida et al. 1996; McLaurin & Steel 2000; 2007; Willis 2000; Horton et al. 2004; Kirschbaum & Hettinger 2004; Aschoff & Steel 2011a; 2011b; Kirschbaum & Spear 2012) and is well-constrained in terms of stratigraphic architecture owing to its high-quality and laterally continuous exposure, which permits relatively high-resolution correlations between alluvial strata and coeval shoreline deposits (e.g. Kirschbaum & Hettinger 2004; Aschoff & Steel 2011a; 2011b). Within the Mesaverde Group different hierarchical orders of depositional sequences have been tentatively established (Yoshida et al. 1996; Willis 2000; Kirschbaum & Hettinger 2004; Aschoff & Steel 2011b). The Neslen Formation has been recognized as being incorporated within a 3rd-order depositional sequence (Yoshida et al. 1996; McLaurin & Steel 2000) and this broadly corresponds to the highstand systems tract of a more generally 'high-order' sequence, the origin of which has been interpreted as being tectonically driven (Willis 2000). Higher frequency 4^{th}- and 5^{th}-order sequences have been distinguished within the Neslen Formation, and their origin has been attributed to forcing by allogenic factors (Aschoff & Steel 2011b).

The Neslen Formation is Campanian in age, attains a maximum thickness of ~120 meters around the Utah-Colorado border, but thins westward to less than 40 m. It consists of interbedded fine- to medium-grained sandstones, mudstones and coals, overall interpreted as having been deposited in a lower coastal-plain setting (cf. Pitman et al. 1987; Kirschbaum & Hettinger 2004; Aschoff & Steel 2011b; Shiers et al. 2014; Olariu et al. 2015). Some of the sandbodies present in the Neslen Formation have been previously interpreted as the product of infill of distributary channels (Kirschbaum & Hettinger 2004; Aschoff & Steel 2011b). Other sandbodies have been interpreted as variably paralic to shallow-marine in origin; one of these sheet-like bodies, named Thompson Canyon Sandstone Bed (hereafter referred to as TCSB), has continuous exposure throughout the study area and is interpreted as having been deposited in an estuarine, sand spit or shoreface setting (Kirschbaum & Hettinger 2004). The traceability of the TCSB throughout the study area and the indication of shoreline contiguity it provides make it an ideal stratigraphic reference for observations reported in this work. Fine-grained and organic lithologies are interpreted as having been deposited in estuarine or lagoonal settings and in delta-plain mires or marshes (Kirschbaum & Hettinger 2004; Aschoff & Steel 2011b). Importantly, the Neslen Formation, especially in its lower to middle part, is known to include evidence of tidal influence recorded in a variety of forms in the deposits, commonly in combination with indicators of brackish-water conditions (Kirschbaum & Hettinger 2004; Aschoff & Steel 2011b; Steel et al. 2012; Olariu et al. 2015). A microtidal regime is suggested to have been active along the western coast of the seaway, at peak regression during the Campanian, based on numerical experiments (Ericksen & Slingerland 1990; see also Slater 1985).

These characteristics make parts of the Neslen Formation ideal for the scope of this study; this succession permits the characterization of distributary channel fills that were apparently influenced by tidal processes in a lower coastal-plain environment.

Principal study localities were selected in the region of Crescent Canyon, about 30 km east of Green River (Fig. 2b). The outcrops are aligned along a transect that is at high angle with the overall depositional dip. Within this canyon, the studied interval has been walked out along the length of its exposure. Results from the work at Crescent Canyon are complemented by additional observations from Tusher Cayon, the Sagers Canyon area and Westwater Canyon (Fig. 2a).

METHODS

The object of the field investigation involved the recognition of sedimentary bodies considered representative of the preserved infill of distributary channels associated with lower coastal plains. Field study focused on sedimentary bodies that are interpretable as the product of deposition in low-sinuosity distributary channels, for which the potential influence of backwater conditions and water-surface drawdown is particularly investigated. However, in addition, other types of channelized bodies were also observed and descriptions of these bodies are briefly reported in the results below.

The necessity to consider a given type of sedimentary unit and a specific environmental setting meant that the stratigraphic interval of interest within the Neslen Formation at Crescent Canyon was narrowed to a section from the middle of the formation, having its base 11 meters below the TCSB base and its top 27 meters above the TCSB top.

Channel architectural elements selected for detailed study were characterized at outcrop. The stratigraphic position of each channel body was mapped as the vertical distance between the top of the channel body and either the base or the top of the TCSB, depending on occurrence of the element below or above the TCSB, respectively. The internal and external geometries of the channel elements and associated bounding surfaces were captured by means of architectural sketches and photographs. The lithologies of the bodies were described and categorized into lithofacies classes based on sediment texture and structure, and interpreted in terms of depositional or post-depositional processes; the ichnology was also described, and trace fossils attributed to ichnogenera where possible; representative vertical profiles were logged for each body. Paleocurrent readings (N = 240) were determined from the dip direction of sets of cross-stratification and cross-lamination, as well as from imbrication of pebbles and cobbles. All paleocurrent observations were classified by quality, by information on their position in the body, and by relationship to associated lithofacies. Average flow directions were used to reconstruct (trigonometrically) real widths of channel elements from outcropping bodies cut at an angle to the reconstructed cross-stream direction. Spatial relationships shown by the channel elements with surrounding sedimentary units, in both the vertical and horizontal directions, were described. Outcrops were walked out to ensure lateral correlations.

CHANNEL-ELEMENT TYPES

In the region of study of the Neslen Formation, three types of channel bodies are recognized in terms of lithology and formative-channel dynamics: (i) laterally accreting sand-prone channel elements; (ii) laterally accreting heterolithic channel elements; (iii) ribbon aggradational sand-prone channel elements. A brief description is given here for the first two types, whereas ribbon channel elements are treated in greater detail in the following sections.

Laterally accreting sand-prone channel elements

Description.—The observed examples (cf. Fig. 4a, 4c) typically have a thickness of 3 to 5 meters at their axis, and widths that vary from 90 to 500 m, showing variable but relatively limited width-to-thickness aspect ratios (up to at least 20, as based on apparent and partially exposed widths). These elements are composed of barform lithosomes that occur either amalgamated in multilateral bodies or as single storys; vertical stacking that forms multi-story bodies is locally seen. Internally, these elements are formed by dm- to m-scale clinothem accretion increments, commonly with sigmoidal cross-sectional geometries in the cross-stream direction, with accretion surfaces dipping at 6 to 20°, and resting on an erosional base. Accretion surfaces display convex-upward geometries in the downstream direction; erosional reactivation surfaces that roughly parallel accretion surfaces are locally seen. Geometries often appear distorted by differential compaction. The dominantly lateral direction of accretion is demonstrated by the high angle between paleo-flow direction (as indicated by cross-strata and cross-laminae dip directions, which also testify to essentially seaward-directed unidirectional flow) and accretion direction in barform elements.

The facies organization of these bodies is dominated by thickly bedded moderately to well-sorted fine-grained sandstones, with subordinate medium-grained sandstones, pebbly intraclast conglomerates, and siltstones. The sandy lithologies often show cross-stratified, cross-laminated or massive to faintly laminated structure. Low-angle (<15°) cross stratification is seen, locally in the form of wide shallow festoons. In places, organic and/or muddy drapes are seen along cross-strata foresets and bottomsets. Lags of pebbly intraclasts are observed at various levels along vertical sections. Thin-bedded, laminated silty deposits are not uncommon but are only present in volumetrically minor proportion (less than 1% as determined from logged thicknesses), preferentially toward the top of the bodies, commonly capping fining-upward trends. Where three-dimensional exposure is available, no particular horizontal trend is observed with regards to the fractionation of sand and mud at different barform positions. *Skolithos* trace fossils were identified in one of these bodies, just below the study interval. In the examined examples, clear evidence of tidal modulation (cf. Shanley et al. 1992; Martinius & Gowland 2011) was not seen.

These bodies are encased in mudstones, organic mudstones or coals; some of the deposits in which these bodies are hosted are rich in thin (prevalently cm-thick) sandstone beds.

Interpretation.—Both the external geometries of these bodies and the lithofacies types which compose them internally are compatible with deposition by fluvial processes in a channel setting (cf. facies models in Colombera et al. 2013). Consistent evidence of unidirectional flow is indicative of lack of flow reversal; evidence of tidally driven flow retardation is not seen in the studied examples. The observed relationships between internal (accretion) geometries and inferred paleocurrent directions are suggestive of dominance of lateral accretion as typically observed on point bars (Bridge 2006). Observations on the deposits in which these bodies are encased are consistent with previous interpretations of coastal plains that are locally rich in organics and transitional to paralic deposits (Kirschbaum & Hettinger 2004; Aschoff & Steel 2011b). Part of the mud-prone deposits in which these units are encased are thought to represent abandoned-channel fills, though well-defined cuts of the supposed outer bank could not be traced reliably. This channel-element type is therefore interpretable as the product of deposition by laterally migrating coastal-plain fluvial channels.

Laterally accreting heterolithic channel elements

Description.—Observed examples (cf. Fig. 4b, 4e) typically have a thickness of 2 to 5 meters at their axis, and measured widths that vary from 50 to 300 m, showing variable width-to-thickness aspect ratios (10 to 60, as based on apparent and partially exposed widths). These elements are also made of barform lithosomes that occur amalgamated in multilateral bodies or as single storys. The barforms are formed by cm- to dm-scale bedding with apparent tabular- or wedge-shaped cross-sectional geometries in the cross-stream direction, dipping at 4 to 25°, and resting on an erosional base, often displaying tangential terminations. Geometries may appear distorted by differential compaction. The dominantly lateral direction of accretion is demonstrated by the high angle between paleo-flow directions and accretion direction in barform elements (e.g. Fig. 4e). Rare occurrences of opposing dip directions are displayed by ripple foresets.

The facies organization of these bodies is dominated by moderately to well-sorted medium- to very fine-grained sandstones and siltstones, which define thinly to thickly bedded (cm- to dm-scale thickness) heterolithic accumulations. The relative proportion of sandstone and mudstone beds within these units is variable, but sandstones are typically dominant and display dm-scale thickness, whereas mudstone beds are commonly subordinate and show cm-scale thickness. The sandy lithologies are typically cross-laminated or cross-stratified, often at low angle. Cross-laminated sandstones are also combined with fine-grained deposits into flaser, wavy, or lenticular bedding. Single and double drapes on ripple foresets and horizontal laminae are also seen, though they do not display evident rhythmicity. Heterolithic packages may also assume the form of sub-cm-scale sandstone-siltstone pin-stripe couplets (cf. Rahmani 1988) that grade into apparently massive siltstone to claystone beds, and on close examination exhibit subtle modulated rhythmicity in the thickness of the sandy vs. silty components (Fig. 4d). Single and double drapes are observed on ripple foresets. Lags of pebbly intraclasts are present in places at the base. *Teredolites*, *Arenicolites*, *Diplocraterion*, and *Rhizocorallium* trace fossils were identified in bodies of this type.

These bodies are typically encased in organic mudstones or coals.

Interpretation.—This channel-element type is interpreted as the product of deposition by laterally migrating coastal-plain channels, the fill of which accumulated in response to important and rhythmical variations in environmental energy, at rare times associated with in-channel flow reversal, and possibly salinity changes, likely under the forcing of tidal processes within the fluvial-tidal transition zone (cf. Dalrymple & Choi 2007; Van den Berg et al. 2007). Part of the mud-prone deposits in which these units are encased are thought to represent abandoned-channel fills, though well-defined cuts of the supposed outer bank could not be traced reliably.

LITHOFACIES AND ARCHITECTURE OF RIBBON CHANNEL FILLS

The lithological characters of the ribbon aggradational sand-prone channel elements can be synthesized into 12 lithofacies classes based on sediment texture and structure. The lithofacies characters, summarized in Table 1, are interpreted in terms of depositional or post-depositional processes. The lithofacies classes are coded following the scheme proposed by Miall (1996), with some additional categories (cf. Colombera et al., 2013). Representative photographs of the lithofacies types are reported in Figure 5 and 6.

The architectural characteristics of the seven ribbon channel bodies that have been studied in detail are summarized in this section via an account of the external geometry of the bodies, of the geometry and nature of their internal bounding surfaces, of their specific lithofacies organization, and of the observed relationships with adjacent bodies. Peculiar features of each body are highlighted in the case-by-case summary provided in the supplemental material. Channel-body examples and paleocurrent data are presented in Figure 7, 8, and 9; representative vertical sections are included in Figure 10 and 11; data on channel-body geometry and lithofacies are reported in Table 2.

Channel bodies of this type are characterized by concave-up erosional bases and flat horizontal tops; they display values of maximum axial thickness ranging from 4.3 m to 6.5 m, and values of true cross-stream width ranging from 35 m to ~100 m (Table 2). The width-to-thickness aspect ratio of these bodies ranges between 7.4 and 18.2, and averages 12.2.

Internally, these bodies are characterized by accretion increments that consist of beds bounded by sharp surfaces or gradational bases, and typically display planar tabular or scoop-shaped geometries. These beds are arranged in bedsets with erosional bases that display limited (cm-scale) relief (bodies A, D, F), or in storys (*sensu* Friend et al. 1979) with lenticular geometry – as seen in 2D outcrop exposures – and erosional bases that display more pronounced (dm- to m-scale) relief (bodies C, E, F, G; Fig. 9b). Beds and tabular bedsets are vertically stacked; scoop-shaped scour fills and storys demonstrate laterally offset axes and vertically offset tops. Although gently inclined accretion surfaces are locally seen, bedsets are typically planar-tabular; bedsets that bear evidence of lateral or downstream accretion are lacking within these bodies. This architecture testifies to the overall aggradational nature of these channel fills. Scour surfaces that form the base of bed-scale scour fills and storys appear to be concentrated towards the base of some of the channel bodies (bodies C, E, F), and tend to be overlain by intraclast-bearing sandstones, which are commonly massive, and conglomeratic lags.

Potential, though highly uncertain (see Table 1), cryptic indicators of tidal modulation and indicators of brackish salinity are locally seen across the vertical profile of the ribbon channel bodies. The first appearance of organic – locally also muddy – drapes on foresets and bottomsets is recorded at different heights above the channel-body bases at their axes, and in association with different lithofacies (Sp at 2.4 above base of body A, Fig. 7a; St at 3.5 in body C, Fig. 7c-d; Sl near base of body E, Fig. 8, in which abundance increases markedly in Sl/Sc deposits at ca. 2 m above base; St near base of body F, Fig. 9a-b; Sl at 0.8 m in body G, Fig. 9d). Sigmoidal cross stratification is seen within the uppermost 1.5 m of channel-body E. Rare instances of herringbone cross-bedding or cross lamination are observed in two channel bodies, in both cases near to the channel-body top (> 4 m above base of body A; ca. 4 m above base of body E). Paleo-landward-directed (i.e. broadly westerly and opposite to dominant readings) paleocurrent directions are also recorded in the uppermost portions of body C and in the lower third of body D. Occurrences of *Skolithos* trace fossils are seen within 2 m of the base of body C and near the top of body E; *Ophiomorpha* is seen at ca. 2 m of the base of body F.

Where exposed, channel-body margins are erosional: they demonstrate relatively steep (up to at least 35°) cut-banks incised, at the channel-top height, into organic mudstones. These channel bodies are contained within muddy or coaly packages, which are at times associated with tabular sandstone bodies of probable paralic or shallow-marine origin.

These sandstone bodies are characterized by heterolithic planar and wavy horizontal lamination, herringbone-ripple cross-lamination, wave-ripple cross-lamination (locally mud-draped or forming wavy bedding), massive structure associated with abundant bioturbation, and by the occurrence of horizontal and vertical burrows (including *Skolithos*; very common), u-shaped burrows (*Arenicolites*) and dm-wide pillow-shaped flat-topped structures interpreted as resting traces.

A case-by-case summary of the channel bodies is provided in the supplemental material.

DISCUSSION

Assessment of the Potential Influence of Backwater Processes on Channel Fills

Aggradational ribbon channel fills are common features of fluvial systems of different types. For example, a tendency to interpret architectural elements of this type in terms of channels in alluvial settings with anastomosing channel pattern was in vogue in the 1980s and 1990s (cf. Makaske 2001, and references therein), largely on the basis of the inferred channel stability, and even though evidence of the existence of multi-channel networks cannot be confidently obtained in outcrop studies. The possibility that the ribbon channel fills described here were deposited in a purely alluvial setting is discarded on the basis of the interpreted paralic origin of deposits with which these channel fills are locally interbedded and that form downdip correlative units. We therefore exclude that the ribbon channel fills could represent the preserved product of alluvial trunk channels that fed more sinuous and mobile coastal channels, to which the lateral-accretion units would be associated. Building upon pre-existing work on the sedimentology and physical stratigraphy of the Neslen Formation, and considering the observations presented above, the ribbon channel bodies described in this work are interpretable as the infill of distributary channels in a coastal plain setting. It is therefore possible to use the sedimentological data presented to assess the occurrence of the features on which criteria have been proposed in this study as potentially diagnostic of the influence of backwater hydraulics on the facies organization of distributary channel fills. It is important to note that the ribbon channel bodies examined here – which occur at multiple levels within the stratigraphy – are almost certainly associated with different distances from their contemporaneous shoreline (cf. Kirschbaum & Hettinger 2004; Aschoff & Steel 2011b; Kirschbaum & Spear 2012): not only could this explain any intrinsic variability in their degree of approximation to the proposed model, but it could also offer the opportunity to compare bodies associated with different likelihood of undergoing and recording the predicted effects of water-surface drawdown at high flow stage. Channel bodies E and F may be the most likely to record the process of water-surface drawdown, given their stratigraphic proximity to the TCSB and the interpreted origin of the deposits with which they are interbedded, provided that discharge variability and system gradient, which control the length of the drawdown zone, were kept constant during the time embodied in the stratigraphic interval, and assuming that no significant base-level fall is recorded in the interval in which they are contained (see below).

The narrow width of these channel fills and the lack of associated accretionary banks reflect formative-channel planform stability. This is in agreement with observations of diminished channel mobility in the backwater zone of some modern rivers (Mississippi and Rhine-Meuse deltas; cf. Kolb 1963; Hudson & Kesel 2000; Gouw & Berendsen 2007; Blum et al. 2013). However, the same characters, as well as the observed reduction in channel mobility near to coasts, may also arise in relation to enhanced bank resistance (Makaske 2001; and references therein), for example. The erosional nature of the channel banks and the lack of co-genetic sand-prone levee or crevasse-splay deposits indicate that the formative channels did not develop elevated channel ridges, nor were they responsible for any significant export of sandy sediment to proximal overbank areas. This is partially in accord with the possibility that streambed deepening and the reduction in water-surface elevation in the zone of drawdown hydrodynamics may inhibit the construction of proximal overbank sandbodies, but it is notable that what is observed is not a reduction in the size, frequency of occurrence and sand content of proximal floodplain units, as postulated. Rather, levee or crevasse-splay sandstones associated with the ribbon channel bodies are lacking altogether. To account for the lack of co-genetic overbank sandstones, the ribbon channel bodies could be interpreted as representing the infill of incisional depressions, possibly representing relatively small tributaries, connected with a minor relative sea-level fall. However, the characteristics of the bounding surfaces associated with these bodies do not particularly support an interpretation that invokes the development of paleotopography related to a base-level fall, in particular in consideration of the geometry of the exposed preserved cut-banks and their terminations at the height of the sand-filled channel-body tops: no contrast in texture, structure or color above the mudstones incised by the channel bodies is observed that would suggest a continuation of surfaces with paleotopographic significance, i.e. that may represent low-relief interfluves. In addition, the facies organization of the ribbon channel fills does not seem to be reconcilable with the possibility that these bodies represent the backfilling of small tributaries that lacked upstream connection with a perennially discharging fluvial system. The limited vertical and lateral extent of these bodies does not particularly support an incised-valley interpretation. Nevertheless, the value of the characteristics of proximal overbank deposits discussed here as possible criteria for the recognition of drawdown influence is entirely speculative: although there has been consideration of the potential feedback of overbank flooding on backwater processes in the channels (Lamb et al. 2012), the importance of backwater hydrodynamics for overbank sedimentation still needs the attention of modern and experimental studies.

Consideration of process interpretations for the internal organization of the ribbon channel bodies is required for the following reasons: (i) to assess whether the processes associated with river-flood intensification by water-surface drawdown are possibly recorded in some of these bodies; (ii) to determine what type of sedimentological imprint is left by these processes in distributary channel fills and how this signature contrasts with characters of low-stage backwater deposition (interflood). A fundamental hypothesis of this study is that the process of cut-and-fill associated with events of water-surface drawdown could be embodied by base-of-channel deposits that would generally thicken toward the shoreline (i.e. to maximum drawdown, where scour depth is largest; Lamb et al. 2012) and possibly record suppression of tidal signature by the decelerating flow during flood recession, subsequent to flood-peak scouring. This possibility would be realistic assuming rapid streambed adjustment, which has been observed in modern distributary channels (cf. Meckel 1975). However, experiments based on binary discharge conditions suggest that the scoured topography generated at high flow is progressively – and not instantaneously – filled during low-flow conditions (Chatanantavet & Lamb 2014); it is therefore possible that only very limited portions of the scours generated during conditions of drawdown hydrodynamics aggrade during flood recession. Furthermore, physical and numerical experiments indicate that the infill of the scours would take place at progressively downstream positions along a given reach (Chatanantavet & Lamb 2014); in a context of progressively decreasing discharge and water-flow velocity, this experimental fact is counter to the argument that fluvial-dominated deposits at the base of the channel fills should thicken toward the shoreline, because it suggests instead that the infill of base-of-channel drawdown-related scours occurs at the upstream end of the drawdown zone during the moments successive to peak flood (i.e. when the obliteration of tidal modulation is maximum), and only propagates downstream as the flood recesses. According to the working hypothesis, whether the upper part of a distributary channel fill contains evidence of tidal influence is principally a function of the state of infilling of the channel and the timing of flood-driven avulsion upstream, together with the relative strength of fluvial and tidal currents. However, on the basis of experimental results (Chatanantavet & Lamb 2014), the infill of distributary channels in the zone of water-surface drawdown can be anticipated to occur at low stage (backwater); thus, if evidence of tidal influence on fluvial processes is preserved from erosion, then it is expected to be present across most of the vertical profile of the infill of a distributary channel. If the occurrence of organic drapes was by itself a reliable indicator of tidal modulation, the basal portion of ribbon channel body that remained unaffected by tidal processes would be limited to the lowermost 2.4 m for body A, 1.3 m for body E, and 0.8 m for body G (Fig. 10 and 11). For the channel bodies A, C and possibly E, landward-directed paleocurrent directions are observed only in their upper portions; this would be in agreement with the hypothesis of current reversal being overridden by accelerated flow during post-drawdown aggradation, meaning that evidence for landward-directed flow is only recorded during low-stage deposition. Landward-directed paleocurrents are instead present in the middle of the infill of body F and from the lower portions of body D. The vertical position of trace fossils that indicate brackish-water conditions provide a maximum measure of the potential thickness of the portion of channel fill that may have been filled during the supposed episodes of post-drawdown flood recession, and thus evidence of whether accumulation may have taken place in the zone of water-surface drawdown if the proposed hypotheses were true. In body C, the occurrence of *Skolithos* likely indicates brackish-water suspension feeders within 3.5 m from the channel top, and may suggest that only the lowermost 2 m there, or only the lowermost 3 m of a story at outcrop 2, may record drawdown-event incision and flood-recession fill (cf. Fig. 10). Similarly, the occurrence of *Ophiomorpha* in body F suggests that a colonization window was briefly open after deposition of the basal 1.7 m, indicating that only the scours occurring below this level may record infill during receding drawdown (cf. Fig. 11). On the basis of these considerations, channel bodies C and E appear to be the ones that match most closely with the sedimentary architecture predicted by the model, whereby a basal portion of the channel fill was influenced by drawdown such that it reveals no evidence of tidal influence. This particular hypothesis of drawdown-controlled facies organization of distributary infill offers an autogenic explanation for the recognition of purely fluvial deposits lying at the base of tidally influenced fluvial distributary channel fills (cf. data from: Meckel 1972; Plink-Björklund 2005; Portela et al. 2009). However, even in such cases (channel bodies C and E) where scour fills in the lower portions of the channel bodies possess no unequivocal evidence of tidal influence, it is still not understood whether these basal portions represent thalweg deposits that are less likely to record tidal processes – possibly in landward portions of the tidal-fluvial transition zone –

or actually the hypothesized storys that represent the infill of ephemeral or event-related scours associated with drawdown. Upstream of the drawdown zone, the presence of scour surfaces at multiple vertical positions in the channel bodies can also be explained by changes in channel hydraulic geometry in response to variations in water discharge, for example related to avulsions occurring upstream in the distributary network (cf. Yalin 1992), and it is not clear how these would be distinguished from drawdown-driven scours. For all the studied ribbon channel fills, the lack or scarcity of mud-prone lithofacies and deposits interpretable as the product of fluid-mud accumulation may suggest that deposition took place either upstream or downstream of the turbidity-maximum zone (cf. Dashtgard et al. 2012; La Croix & Dashtgard 2015; for sedimentological criteria applicable to barform deposits). The occurrence of *Ophiomorpha* and/or *Skolithos* trace fossils in some of the channel fills is indicative of likely deposition in a saltwater-influenced reach of the backwater zone, but not suggestive of whether deposition may have taken place downstream of the turbidity-maximum zone. The facies organization of the ribbon channel fills is compatible with what is observed in a number of other outcropping ancient successions that are interpreted as having accumulated in a lower coastal-plain setting (Okolo 1983; Hopkins 1985; Dreyer 1990; Kirschbaum & McCabe 1992; Olsen 1993; Plink-Björklund 2008; among others), and part of which bear a record of tidal influence. In some cases, the observed spatial relationships with mouth-bar, bay or delta-front deposits permit the identification of bodies as the infills of terminal distributary channels near their mouths (Ryer 1981; Okolo 1983; Ryer & Anderson 2004; Olariu et al. 2005; Olariu & Bhattacharya 2006; Rygel & Gibling 2006; Plink-Björklund 2008). In all these cases, instances of sand-prone cut-and-fill aggradational ribbon channel bodies are observed that display the following dominant characteristics: (i) cross-stratified and cross-laminated sandstones, in some cases convoluted; (ii) massive and planar-horizontal stratified sandstones; and (iii) a scarcity or absence of fine-grained sediment. A comparable sand-prone character is also seen in Holocene deposits interpreted as the infills of distributary channels in tide-influenced lower delta plains (cf. Hutchinson et al. 1995; Fielding et al. 2005).

To partly confront the uncertainty that exists as to whether the deposition of the studied ribbon channel bodies took place within the zone of influence of the process of water-surface drawdown, it is useful to consider what the likely length of this zone may have been. The degree to which the studied bodies may have undergone the influence of water-surface drawdown is not just a function of their distance from the shoreline, but also a function of the system gradient and water discharge variability. For the clastic wedge in which deposits of the Neslen Formation are incorporated, it has been suggested that reduced basin gradient may have resulted from the interference between Sevier-style tectonics, Laramide-style tectonics, and dynamic subsidence, as this would offer explanation of the observed extensive transgressions, which are partially attributed to eustatic fluctuations (Aschoff & Steel 2011b). A low-gradient coastal plain would have been characterized by a relatively long backwater zone (Paola & Mohrig 1996; Lamb et al. 2012), which could therefore be characterized by a large difference between the size of backwater length (i.e. transitional zone at low flow) and the size of drawdown length (i.e. transitional zone at high flow). In addition, the Neslen Formation is known to have been deposited during greenhouse climatic conditions (Huber et al. 2002), considered favorable to high-precipitation events that could drive high-discharge floods, and more generally to significant discharge variability that could have led to a marked distinction between flood-driven drawdown and low-stage backwater. The paleolatitude at which the Neslen Formation accumulated is estimated to have been between 44° and 47° North (Miller et al. 2013), thus within a latitudinal belt for which high and seasonal precipitation is inferred on the basis of leaf physiognomy (Wolfe & Upchurch 1987). Also, isotopic evidence exists for the presence of a strong monsoon along the eastern side of the Sevier Orogen during Neslen times (Fricke et al. 2010). Overall, this situation could have been conducive to the development of a marked differentiation between an upper delta plain dominantly subject to flow deceleration and deposition, and a lower delta plain influenced by diminished sediment flux at low flow, and episodically subject to flow acceleration and channel incision at high flow. Due to its dynamic character, the extent of the zone of drawdown varies in relation to the magnitude of the events; as a tentative control, it is valuable to estimate the likely scale of the backwater or drawdown length for the Neslen system. First of all, it is possible to consider estimates of coastal-plain slope as inferred from the gradient of transgressive surfaces traced by Aschoff & Steel (2011b): this returns a gradient of ca. $2.5*10^{-4}$ m/m. This figure carries significant uncertainty, in relation to the fact that the considered surfaces are not representative of the graded profile on land, are based on correlations, and no correction has been applied to account for compaction. Relying on maximum bar thickness (cf. Bhattacharya & Tye 2004) or on cross-strata set thickness (cf. Yalin 1964; Bridge & Tye 2000; Leclair & Bridge 2001) as a mechanism for reconstructing bankfull depth of channels in the Neslen Formation, values of 7.5 m (Lawton 1986) or ranging from 7.2 m to 12.2 m (this study) are derived, respectively. Considering a bankfull depth of 8 m as representative of the scale of the channels in the Neslen system, an indicative backwater length of ~30 km is derived as the ratio between flow depth and gradient (Paola & Mohrig 1996; Blum et al. 2013); however, Lamb et al. (2012) warn that this value should be more properly accounted as an approximation of twice the drawdown length in cases when the normal-flow depth is large relative to the flow depth at the river mouth and under small Froude number conditions, which would therefore yield a representative drawdown length of ~15 km. This seems to suggest that the portion of the depositional system that may have undergone the drawdown process – and more generally backwater effects – had a relatively limited size, but possibly large enough to generate autogenic stratigraphic trends associated with the expected variations in channel behavior, which may have combined with allogenic changes.

In spite of the lack of clear evidence, it is tempting to consider backwater hydraulics as offering a possible explanation for the differences observed between the three types of channel bodies discussed herein (sand-prone laterally accreting bodies, heterolithic laterally accreting bodies, and ribbon sand-prone aggradational bodies). The different architectural types co-occur in the same interval, but it is unclear whether they occur in the same geomorphic context, as deposits of the Neslen Formation represent a low-gradient coastal plain in which modest relative sea-level changes may have shifted the shoreline position significantly. Consideration of backwater effects in modern systems (cf. Blum et al. 2013) suggests that the sand-prone lateral-accreting bodies may represent channels in the upper backwater zone, where they undergo rapid migration and are sand-rich in relation to the backwater enhancement in bedload deposition. Heterolithic lateral-accreting bodies may represent sinuous mobile channels that experienced significant tidal influence, possibly developed closer to the shoreline than the previous type. The ribbon channel bodies could represent the product of laterally stable channels in the most distal reaches of the coastal plain, where distributaries were subject to starvation-related straightening and were more likely to undergo drawdown. However, the spatial distribution and stratigraphic resolution of the data presented here are not sufficient for testing this hypothesis. Also, although increased lateral stability is expected for distributary channels in the lowest reaches of a coastal plain, observations from modern systems demonstrate that variably sinuous meandering distributary channels may occupy the lowermost portions of a delta. The same situation could apply to the distributary channels of the Neslen Formation: it is conceivable that, due to episodic backwater-length shortening, possibly related to hydrological changes, the control exerted by segregation of bedload in the upper backwater zone may have been effective only intermittently.

Recommendations for Future Research

Further work is recommended for the following reasons: (i) to make better use of the growing body of evidence given by numerical and physical experiments and studies of modern systems to develop sedimentological criteria that enable processes connected with backwater hydraulics to be discerned in distributary channel fills from the ancient rock record; (ii) to better assess these criteria through outcrop studies; (ii) to determine whether stratigraphic variations in channel-body architecture could be related to the control exerted by backwater effects. Results would have implications concerning the ability to infer paleo-shoreline movements on the basis of sedimentological and stratigraphic variations. In contrast to the analysis presented here, to achieve the latter two objectives, it is imperative to collate architectural data, on all the different channel-body types, distributed across suitable depositional dip-oriented successions, over distances comparable to the backwater length of each case-study depositional system, and collected in a high-resolution stratigraphic framework. The proposed hypotheses of how in-channel and overbank deposition may be controlled by backwater processes could be better tested if spatially distributed data along shore-normal transects were available, and would need to be considered together with sedimentological criteria that relate facies organization to depositional-dip position in relation of tidal-fluvial interaction (cf. Dalrymple & Choi 2007; Dashtgard et al. 2012; Jablonski & Dalrymple 2016; La Croix & Dashtgard 2015). Additionally, types of field observations that could be gathered in consideration of their possible relations to backwater controls include the following:

- the overall proportion of channel deposits in stratigraphic intervals, which could decrease downstream toward lower backwater reaches due to increased planform stability and reduced likelihood of avulsion (cf. Chatanantavet et al. 2012);
- the relative proportion of channel-fill and barform deposits (cf. Shiers et al. 2014, for preliminary results): both types are likely to co-occur in the backwater zone, as seen in the geomorphology of modern deltas, but their relative proportion may be indicative of backwater influence on channel morphodynamics (cf. Blum et al. 2013);
- the width-to-thickness aspect ratio of channel bodies, which is expected to decrease seaward for both channel fills and point bars, in relation to drawdown-related scouring and reduced lateral mobility, respectively (cf. Fig. 1); variations in architectural styles should be documented to discriminate the potential influence of backwater processes from the effect of flow splitting on channel-body geometry.

Given that the alternation of backwater and drawdown hydrodynamics are inherent to coastal rivers with temporally variable water discharge, evidence of seaward displacement of the landward limit of tidal modulation by seasonal discharge fluctuations may expected to be seen in co-occurring barform deposits that record tidal influence (cf. Sisulak & Dashtgard 2012; Gugliotta et al. 2015; Jablonski & Dalrymple 2016), such as the ones documented in this study.

To assess the possible obliteration of the effects of tidal influence during receding hydrograph limbs that follow water-surface drawdown at high-flow stages, it would be particularly important to carry out further work on systems containing evident tidal indicators, possibly following a multidisciplinary approach that combines sedimentological descriptions with other methods (e.g. analysis of the vertical distribution of palynomorphs or palynofacies assemblages in channel fills; cf. Gastaldo et al. 1996; Czarnecki et al. 2014).

CONCLUSIONS

On the basis of knowledge of Quaternary and modern systems, and in combination with outcomes of numerical modeling, a qualitative model has been proposed to predict the sedimentological characteristics of distributary channels developed in the seaward portions of coastal plains. Such channels are considered to be subject to a reduction in lateral mobility due to diminished sediment flux at low flow and channel incision due to water-surface drawdown at high flow. The model predicts that distributary channel fills may tend to exhibit low width-to-thickness aspect ratio, aggradational accretionary style, one or more basal lithosomes recording high-flow stage incision and possibly infill, and limited occurrence of co-genetic river-fed overbank sandstone elements.

These potential criteria for the recognition of backwater controls have been assessed against field observations of coastal-pain deposits from a restricted stratigraphic interval of the Neslen Formation in eastern Utah, which embodies a suitable environmental setting. Here, several aggradational ribbon channel fills display erosive banks, sand-prone character, occurrence of multiple cross-cutting erosive surfaces toward their base, higher concentration of cryptic tidal indicators in their upper bedsets, and lack genetically related overbank sandstones; these observations are all in agreement with the proposed model. Some of these channel bodies exhibit basal lithosomes that represent scour infill and lack potential tidal indicators. However, the facies organization of the studied bodies does not appear to be in clear agreement with what was hypothesized. In particular, uncertainty exists as to whether particular sedimentary features (organic drapes, sigmoidal cross-strata) indicate tidal flow modulation, and as to how the facies organization of distributary channel fills should be controlled by non-uniform flow in the backwater zone and by its interaction with tidal processes. In addition, it is unclear whether these channel fills developed within the zone of drawdown influence. Hence, additional evidence is required in support of the importance of backwater processes in controlling the architecture of distributary channel bodies in the hypothesized ways.

Laterally accreting barforms were also recognized in the same stratigraphic interval, and their sedimentological characteristics can be tentatively related to a variable interplay between normal alluvial processes, tidal processes, and backwater effects at low-flow stage. Particularly, these barforms may represent the typical product of channel deposition in the upper to middle reaches of the backwater zone of the Neslen rivers, where channel mobility is expected to be typically high and marine influences increase down-system in a manner that controls barform lithology, as reflected in the sand-prone and heterolithic bar types.

Further work is needed to develop and test criteria for the identification of backwater hydraulics in the rock record. Recognizing a systematic distribution of architectural types of channel bodies as a function of backwater influence may have implications for the refinement of sequence stratigraphic models accounting for the evolution of progradational or retrogradational clastic coastal systems.

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TABLE AND FIGURE CAPTIONS

Table 1: Summary of lithofacies types recognized within the ribbon channel bodies of the Neslen Formation in the Crescent Canyon area.

Table 2: Summary of stratigraphic position, dimensions and lithofacies types for the studied channel bodies. Stratigraphic positions are reported as below the base (negative values) or above the top (positive values) of the Thompson Canyon Sandstone Bed (TCSB).

Figure 1: A) Ideal model for the geometric change expected for channel bodies across the transition from alluvial plain to lower delta-plain, and comparison with data from the Lower Castlegate Formation (Petter 2010) and the Mississippi River (Hudson & Kesel 2000; Blum et al. 2013). B) Comparison of the ideal channel behaviors expected at high discharge for alluvial plains and at the shoreline. Distributary channels in the zone of drawdown may approximate to some degree the latter end-member behavior.

Figure 2: A) Map placing the Book Cliffs (Utah and Colorado, USA) in the Western Interior Seaway context, and showing the position of the Crescent Canyon study area and of other visited locations. B) Location of outcrops in Crescent Canyon numbered according to the numeration followed in the text (USGS 2011 US Topo 7.5-minute map, Floy Canyon South, UT; contour interval 40 feet).

Figure 3: Stratigraphy of part of the Mesaverde Group and overlying strata in the Book Cliffs region from Price (UT) to Grand Hogback (CO). The study interval investigated in the Crescent Canyon area (see Fig. 1) is highlighted. Modified after Kirschbaum & Hettinger (2004).

Figure 4: Examples of sand-prone and heterolithic laterally accreting channel bodies from the lower Neslen Fm., seen in the Crescent Butte/Canyon area. A) Sand-prone lateral accreting barform. B) Two amalgamated heterolithic laterally accreting barforms. C) Closer view of the barform in Figure 4a, at a different location; in the upper portion of the bar, a cut-and-fill sedimentary body that may represent a (chute?) channel fill or a scour fill is labeled as ‘CH?’. D) Close view of a mud-prone bed from a heterolithic laterally accreting barform depicted in Figure 4b; lens cap is 5 cm in diameter. E) Heterolithic laterally accreting barform and relative rose diagrams of accretion direction (dip direction of accretion surfaces) and paleocurrent direction. LA labels identify the laterally accreting barforms; traces highlight the bases of the bodies; white arrows indicate the apparent direction of accretion; the TCSB labels indicate the position of the Thompson Canyon Sandstone Bed.

Figure 5: Field examples of lithofacies types. A) Gh/cm with sandy intraclasts, from lower part of channel body C; lens cap is 5 cm in diameter. B) Gh/cm facies with muddy, variably sideritic, intraclasts, interbedded with Sm deposits, from lower part of channel body C; pen in the red ellipse is 15 cm long. C) Sm/s deposits: massive sandstones with pebbly sideritic muddy intraclasts infilling cross-cutting scoop-shaped scour fills, from lower part of channel body C; yardstick is 20 cm long. D) Sh deposits cut by scour fills of massive sandstone with muddy pebble clusters, from lower part of channel body C; yardstick is 20 cm long. E) Sl facies; paleoflow to the right; pen is 15 cm long. F) Sl facies with organic drapes, from middle part of channel body E. G) St facies with variably sideritic muddy intraclasts, from lower part of channel body C; lens cap is 5 cm in diameter. H) St facies with organic drapes, from upper part of channel body C; lens cap is 5 cm in diameter. See the text for further explanation.

Figure 6: Field examples of lithofacies types. A) Sp facies with local concentrations of sideritic muddy intraclasts, paleoflow to the right; pen in the red ellipse is 15 cm long. B) Su facies with transitional base, overlying Sm deposits; pen in the red ellipse is 15 cm long. C) Sr facies with trough ripple cross-lamination; lens cap is 5 cm in diameter. D) Sd facies with

convolution; yardstick in the red ellipse is 20 cm long. E) Sc facies, from channel body C; paleoflow to the right and into the page; lens cap is 5 cm in diameter. F) Sσ facies, from upper part of channel body E; paleoflow to the left and out of the page; yardstick is 20 cm long. G) erosive (reactivation?) surfaces associated with foresets of facies Sσ; yardstick is 20 cm long. See the text for further explanation.

Figure 7: A) Ribbon channel bodies A and B exposed at high angle with their axes (i.e. near to perpendicular to paleoflow), at outcrop 1. B) Ribbon channel body A exposed along its axis just NW of outcrop 7, and rose diagram of paleocurrent directions, including readings from outcrop 1. C) Ribbon channel body C exposed both at high angle with its axis (left-hand side) and along it (right-hand side) at outcrop 2. D) Left-hand portion of ribbon channel body C exposed at outcrop 3. E) Rose diagram of paleocurrent directions for channel body C, including readings from outcrops 2 and 3. F) Ribbon channel body D exposed at high angle with its axis at outcrop 4, and rose diagrams of paleocurrent directions for the channel body and the lenticular sandstones at its base. The TCSB labels indicate the position of the Thompson Canyon Sandstone Bed; traces highlight the bases and tops of the bodies. see Fig. 1b for outcrop location.

Figure 8: A) Ribbon channel body E exposed normal to its axis (i.e. perpendicular to paleoflow) at outcrop 5; the channel-body outline shows the channel downcutting into a tabular paralic to shallow-marine sandstone body; the portion of outcrop in the hatched-line frame is shown in Figure 8B. B) Ribbon channel body E exposed along its axis at outcrop 5. C) Ribbon channel body E exposed normal to its axis at outcrop 6; the portion of outcrop in the hatched-line frame is shown in Figure 8D. D) Ribbon channel body E exposed along its axis at outcrop 6. E) Rose diagrams of paleocurrent directions for the ribbon channel body E at outcrops 5 and 6, represented separately and merged.

Figure 9: A) Ribbon channel body F exposed at high angle with its axis (i.e. near to perpendicular to paleoflow) at outcrop 7. B) Channel body F exposed along its axis (i.e. parallel to paleoflow) at outcrop 7. C) Rose diagrams of paleocurrent directions for the ribbon

channel bodies F and G; D) Ribbon channel body F exposed obliquely to its axis at outcrop 8. Traces highlight the bases and tops of the bodies.

Figure 10: Vertical logged sections for channel bodies A (A) and C (B). Lithofacies codes: Fl = laminated siltstone; Fl/m = laminated to massive siltstone; Gcm = massive conglomerate; Gh = crudely bedded conglomerate; Sc = sandstone with hummocky-like convex-upward bedding; Sd = sandstone with soft-sediment deformation; Sh = planar horizontal bedded sandstone; Sl = low-angle cross-stratified sandstone; Sm = massive sandstone; Sp = planar cross-stratified sandstone; Sr = ripple cross-laminated sandstone; Ss = scour-fill massive sandstone; St = trough cross-stratified sandstone; Su = sandstone with crenulated laminae; Sσ = sandstone with sigmoidal cross-stratification. See text for explanation of lithofacies.

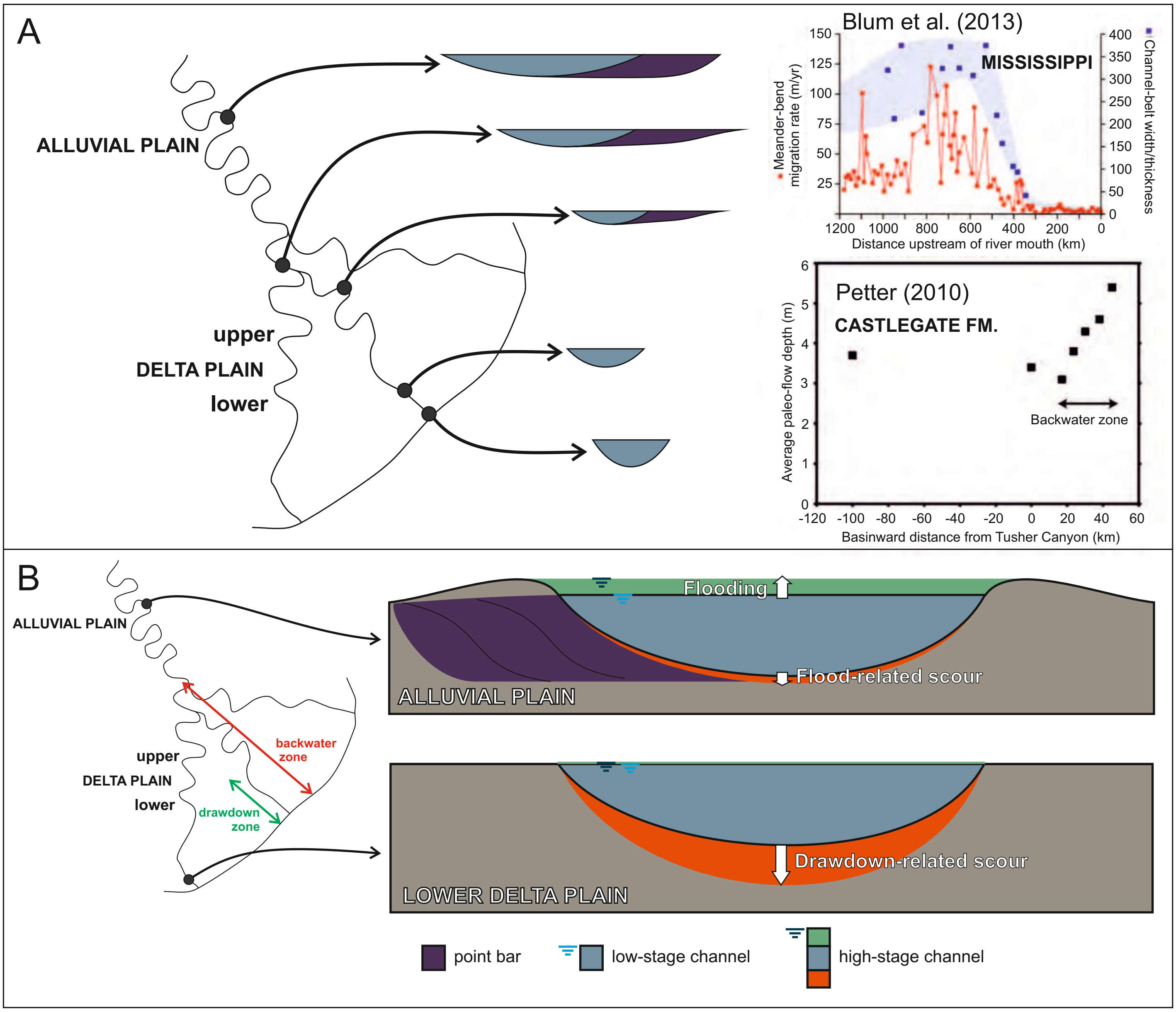
Figure 11: Vertical logged sections for channel bodies D (A), E (B), F (C), and G (D). See Fig. 10 for legend. Lithofacies codes: Fl = laminated siltstone; Sc = sandstone with hummocky-like convex-upward bedding; Sd = sandstone with soft-sediment deformation; Sh = planar horizontal bedded sandstone; Sl = low-angle cross-stratified sandstone; Sm = massive sandstone; Sp = planar cross-stratified sandstone; Sr = ripple cross-laminated sandstone; St = trough cross-stratified sandstone; Sσ = sandstone with sigmoidal cross-stratification. See text for explanation of lithofacies.

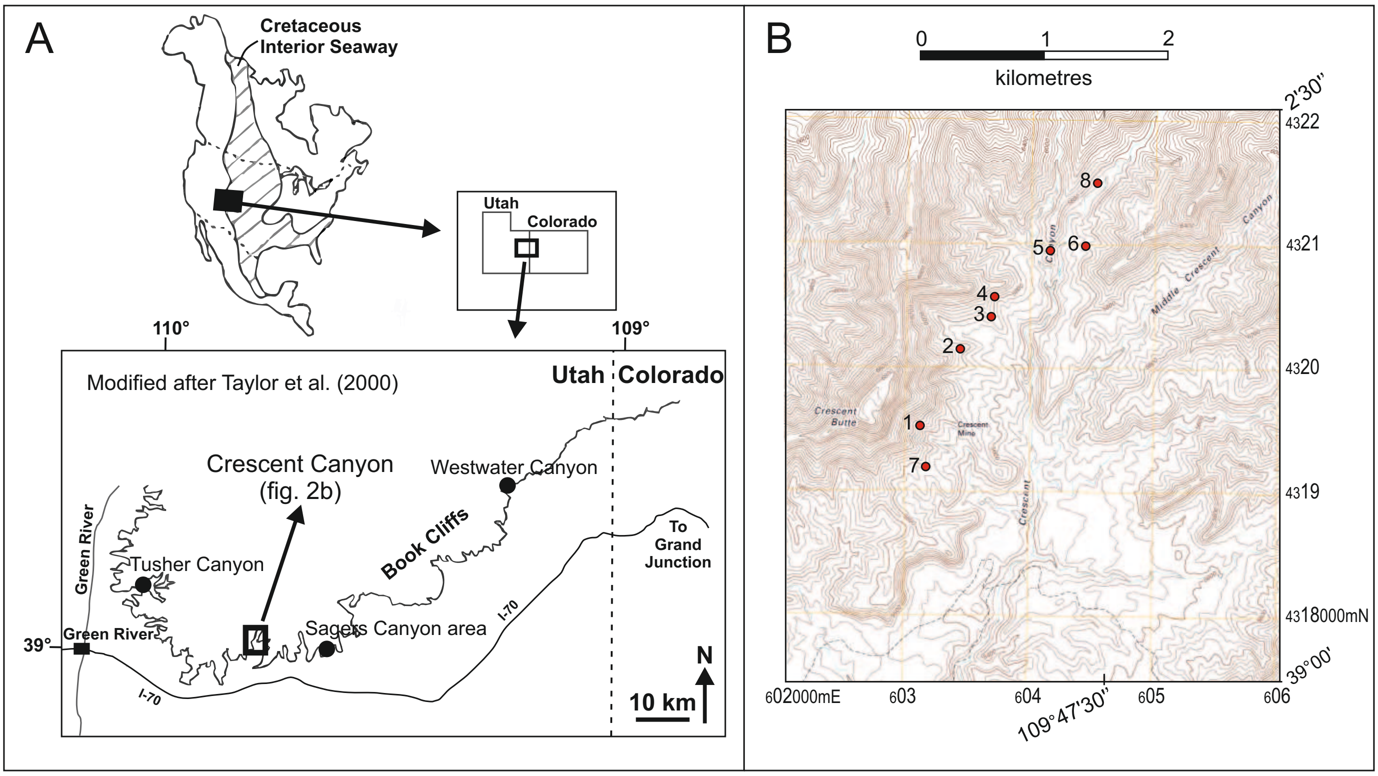
Facies code	Description	Interpretation
Gh/cm	Clast-supported, intraclast conglomerate or breccia, containing granule- to cobble-sized clasts (maximum diameter reaching 20 cm). Massive or crudely horizontally bedded, locally with weak normal grading and unidirectional imbrication. Pebbles and cobbles comprise sandy, muddy or siderite (or sideritized mudstone) intraclasts (angular to well rounded). The interstices between the clasts occluded by fine sand. Gh/cm fills scoop-shaped or lenticular erosive pockets. Upper boundaries often gradational (Fig. 5a, 5b).	Product of deposition – as channel lags or scour fills – of bedload sediment that was mostly reworked within the basin through coastal-plain cannibalization.
Sm/s	Fine- (dominant) to medium-grained (subordinate) sandstone, mostly fairly to moderately sorted, but locally poorly sorted and enriched in fine-grained matrix. Massive or indistinctly laminated, locally with weak normal grading. Concentrations of ironstone (siderite/sideritized mudstone) and muddy pebbles and granules common at the base of beds and as pebble clusters (Fig. 5c, 5d). Potential bioturbation rarely seen. In places, Sm/s demonstrates gradational lateral transitions to faintly laminated sandstones with low-angle concave laminae lined by granule intraclasts. Sm/s typically rests on erosive surfaces and forms dm- to m-thick lenticular or tabular beds. Upper boundaries may be either sharp or gradational. Interbedding with Gh/cm is common.	Often interpreted as the product of rapid deposition of sand from mass flows or hyperconcentrated flows (e.g. originating from the collapse of sandy banks, and locally filling scours; cf. Martin & Turner 1998). Bank collapse may not be the trigger to deposition, as the channel bodies are cut into fine-grained banks, and erosional surfaces and pebble clusters within the deposits suggest pulsating flow. Rather, high-energy conditions associated with high flow stage are envisaged. Where transition to faint lamination is seen, a mechanism of bedform migration cannot be excluded. A post-depositional origin is excluded for occurrences with erosive base and normal grading.
Sh	Fine-grained, fairly to well-sorted, planar horizontally laminated sandstone arranged into dm-thick tabular beds (Fig. 5d). Limited planform exposures impeded recognition of primary current lineation. Sharp boundaries with other lithofacies are seen.	Product of deposition in either the upper or lower flow regime plane-bed field (Collinson et al. 2006).
St	Fine- (dominant) to medium-grained, moderately to well-sorted trough cross-stratified sandstone. Sets are 10 to 50 cm thick, and indicate unidirectional paleocurrents. Ironstone/muddy pebbles and granules commonly distributed along the foresets, locally as clusters (Fig. 5g). Comminuted organic debris commonly forms sub-mm-thick carbonaceous or muddy carbonaceous drapes (‘coffee-ground’ deposits) on foresets, bottomsets or toesets, with apparently non-rhythmical mm- to cm-scale horizontal spacing (Fig. 5h). St displays a sharp base and forms dm- to m-thick lenticular or tabular beds. Upper contacts with other lithofacies can be sharp or gradational (due to convolution).	Product of deposition by migrating three-dimensional dunes. Organic drapes are not reliably interpretable in terms of deposition at temporal energy minima by analogy with tidal structures (cf. Shanley et al. 1992; Plint & Wadsworth 2003; Martinius & Gowland 2011), as they do not tend to assume the form of ‘double’ drapes and do not show rhythmical variations in down-current spacing that could be related to tidal fluctuations, possibly over spring-neap tidal cycles (Van den Berg et al. 2007, and references therein). Deposition of organic drapes on lee sides may occur due to the

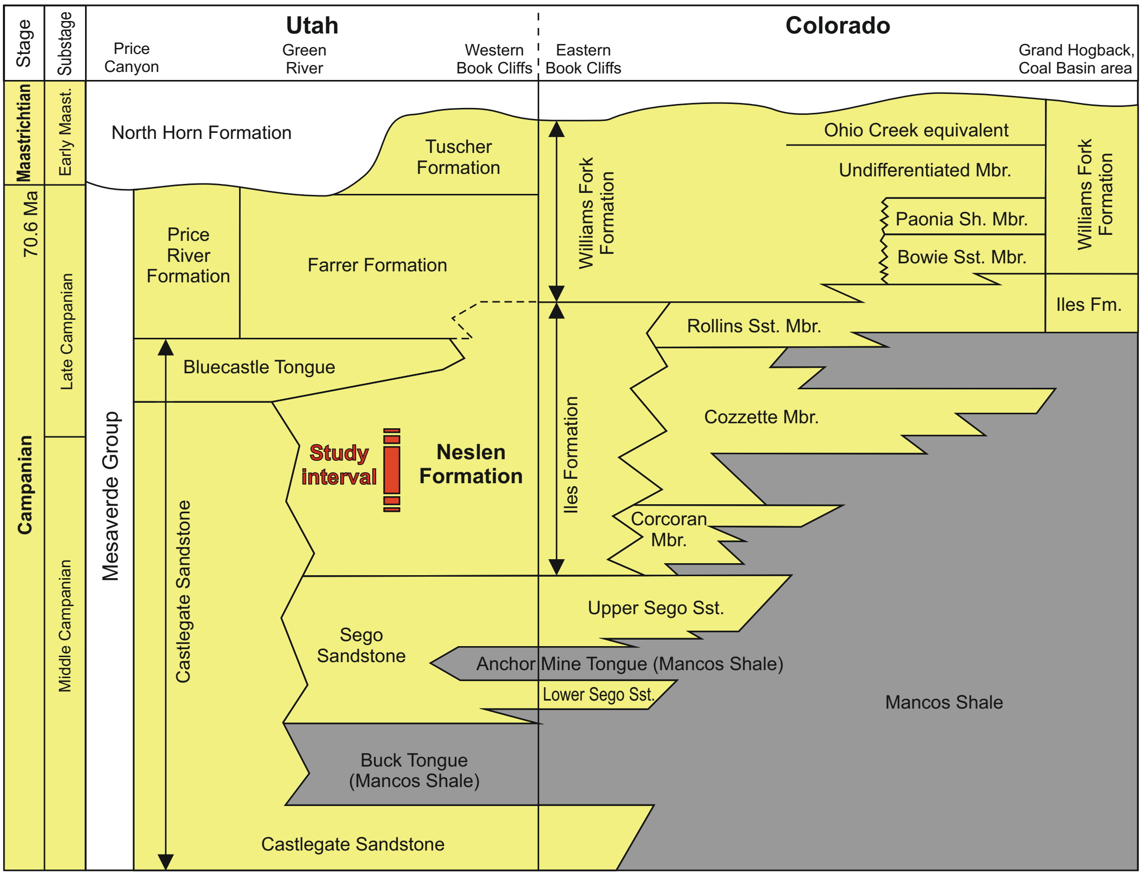
		entrapment of finely macerated phytodetrital material in separation cells (Shanley et al. 1992), even in relation to high-flow stage in inland settings (cf. Alexander et al. 1999).
Sl	Fine- (dominant) to medium-grained, moderately to well-sorted sandstone, with low-angle cross stratification, consisting in planar or shallow festooned concave laminae dipping at less than 15° (Fig. 5e), arranged in cosets with unimodal dip directions. Sets are 10 to 50 cm thick. Ironstone/muddy pebbles and granules are rarely aligned along the foresets. Sub-mm-thick, apparently non-rhythmical carbonaceous (locally muddy) drapes are seen (Fig. 5f). Sl displays sharp or gradational base and forms dm-thick lenticular or tabular beds. Upper contacts can be either sharp or gradational.	Product of deposition either (i) in the upper or lower flow regime plane-bed field on existing low-relief topography, or (ii) by migrating three-dimensional low-relief bedforms. In the latter case, these deposits may have developed under conditions transitional between the lower and upper flow regimes, and may represent bedforms such as washed-out dunes (cf. Fielding 2006). Considerations made for the interpretation of organic drapes in lithofacies St equally apply to these deposits.
Sp	Fine- (dominant) to medium-grained, moderately to well-sorted planar cross-stratified sandstone. Sets are 5 to 50 cm thick and indicate unidirectional paleocurrents. Variably sideritic muddy pebbles and granules locally distributed along the foresets (Fig. 6a). Sub-mm-thick, apparently non-rhythmical carbonaceous (locally muddy) drapes seen on the foresets or the bottomsets/toesets. Sp commonly exhibits a sharp base and forms dm- to m-thick lenticular or tabular beds. Upper contacts can be sharp or gradational (due to convolution).	Product of deposition by migrating two-dimensional dunes (Collinson et al. 2006). Considerations made for the interpretation of organic drapes in lithofacies St also apply to these deposits.
Sr	Fine- (dominant) to very fine-grained, moderately to well-sorted, ripple cross-laminated sandstone; locally faintly laminated. Sets of ripple cross-laminae often arranged into sub-critically climbing dm- to m-thick cosets. Trough cross-lamination locally recognized (Fig. 6c). Toward the top of channel-body A, herringbone cross-lamination is seen. Mud or organic drapes rarely seen on microform lee sides. Sr forms cm- to m-thick lenticular or tabular beds, with sharp or gradational contacts, rarely interbedded with silty deposits. Sharp bases may be lined by granules and pebbles when erosive, and drape the topography when accretionary.	Product of deposition by migrating asymmetric ripples. Climbing cross-lamination indicates high rates of suspension fall out combined with bedload traction. Sinuous-crested ripple forms are demonstrated by the presence of trough-cross lamination. The rare herringbone cross-lamination is indicative of occasional reversal in flow direction (Collinson et al. 2006).
Sσ	Fine-grained, moderately to well-sorted sandstone with sigmoidal cross stratification, which consists of laminae defining foresets that dip up to 24°, pass into tangential toesets and topsets, and show unimodal dip directions (Fig. 6f). Cuts oriented along foreset strike reveal dominantly concave-upward lamina geometries, but low-amplitude convex-upward laminae are also observed in places. Erosional surfaces with limited cm-scale relief and oriented sub-parallel to the laminae bound groups of laminae within sigmoids (Fig. 6g). Rare organic drapes seen on	Product of deposition by unidirectionally migrating mesoforms. The sigmoidal bedding could be indicative of conditions transitional between lower and upper flow regime, under which the migration of humpback dunes could generate this structure (cf. Fielding 2006; and references therein). Evidence of tidal modulation such as rhythmically arranged mud or organic drapes, or variations in foreset angle and

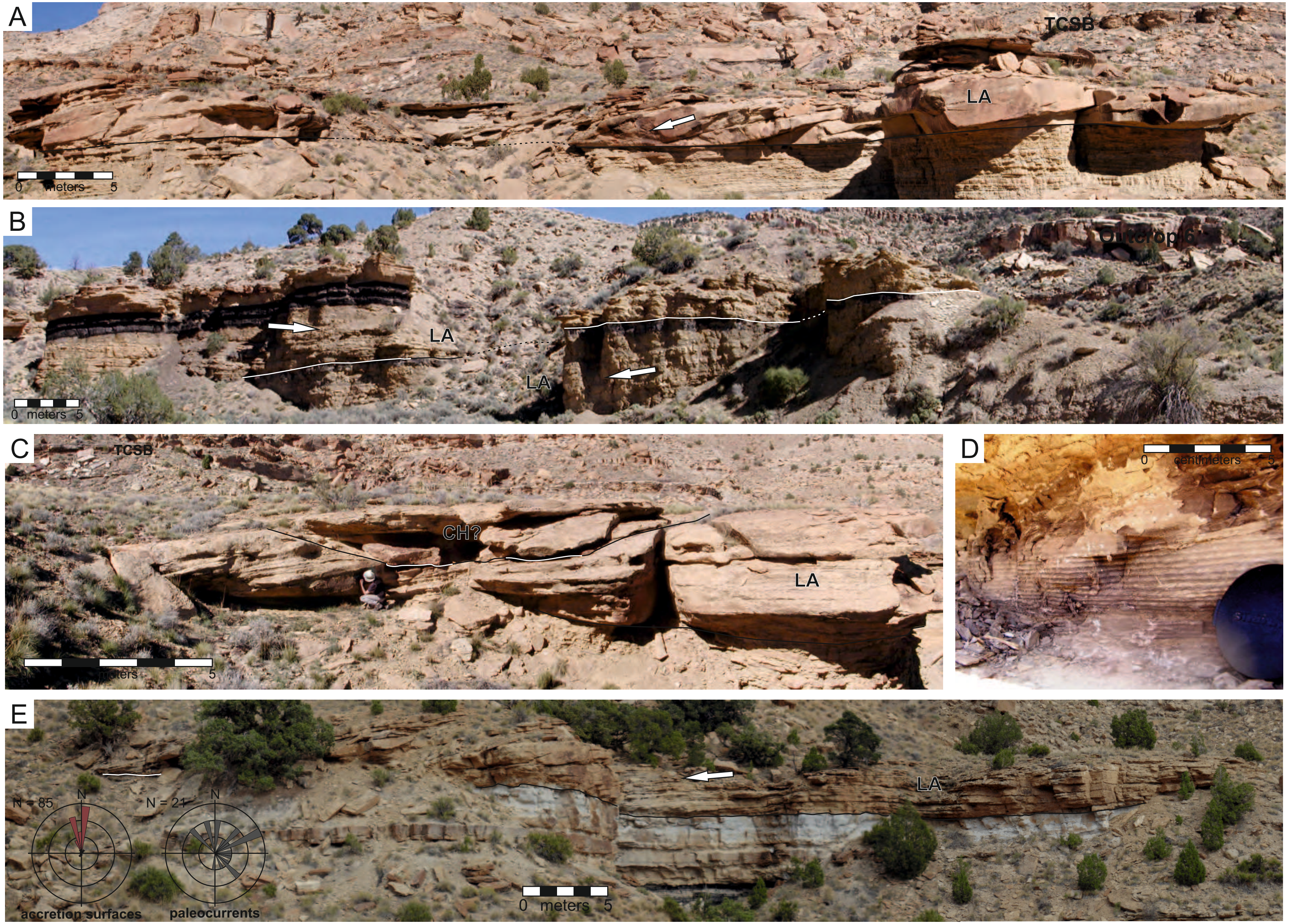
	sigmoidal laminae. This lithofacies was only observed in the uppermost part of channel body E, where it forms a 80-cm thick lenticular bed that displays a gradational base with adjacent lithotypes Sl (below) and St, Sp and Sr (above and laterally).	brinkpoint height (Shanley et al. 1992; Plink-Björklund 2005; Martinius & Gowland 2011; see also: Boersma & Terwindt 1981; Allen & Homewood 1984; Kreisa & Moiola 1986; Uhlir et al. 1988) is not seen. Erosional surfaces within the sets may be interpreted as reactivation surfaces connected to ebb-modulated fluctuations in flow velocity (cf. Nio & Yang 1991; Brettle et al. 2002), but data on their horizontal spacing are lacking due to outcrop orientation.
Sc	Fine-grained, moderately to well-sorted sandstone with convex-upward bedding, which consists of parallel laminae dipping at up to 15º and defining (quasi-)symmetrical hummock-shaped surfaces with dm-scale amplitude and dm- to m-scale wavelength (Fig. 6e). Tangential toesets seen. Laminae are commonly faint, but locally marked by organic drapes and granule-sized intraclasts distributed along their length. Bases of bedsets commonly characterized by flat, sharp, locally erosive surfaces. Sc is observed mostly within the uppermost part of channel body C, in contact with St and Su.	Product of deposition by migrating mesoforms. Analogous deposits recognized within channel sandstone bodies have been interpreted as resulting from deposition either in flow conditions transitional between the lower and the upper flow regime, or in the upper flow regime (cf. Fielding 2006; and references therein); such deposits may, for example, record the migration of symmetrical dunes.
Su	Fine-grained (dominant) to very-fine-grained, moderately to well-sorted sandstone. Though mostly composed of clean sandstone, interlaminated silt is rarely present (body C). The structure consists of undulose lamination, defined in cross-section by parallel, horizontal, symmetrically crenulated laminae with mm-scale amplitude and mm- to cm-scale wavelength. The laminae are mostly faint, but in places marked by lining of granule-sized intraclasts or thin discontinuous mud drapes. Su forms cm- to dm-thick lenticular or tabular beds, with gradational contacts to Sm, Sh, Sl or Sr (Fig. 6b).	Possibly the product of deposition by migrating high-wavelength, low-amplitude ripples. Locally, gradual transitions to Sh and Sl might suggest that the wavy nature may be related to subtle deformation of planar parallel laminae by physical disturbance. Similarity is noted with crenulated deposits described by Sumner et al. (2012) and interpreted as associated with dewatering, but other structures that demonstrate evidence of dewatering are not observed associated with Su.
Sd	Fine- to medium-grained, moderately to well-sorted sandstone with contorted laminae (Fig. 6d), which typically outline antiformal cusps or recumbent folds. Sd forms cm- to m-thick intervals with gradational base or lateral contact with St, Sp, Sl or Sm/s lithofacies.	Product of post-depositional deformation of cross-stratified sandy deposits, probably in response to fluidization (cf. Owen 1995).
Fl/m	Laminated to massive grey siltstone, locally carbonaceous. Where present (bodies A and G), it occurs in the form of cm-thick streaks or dm-thick lenses, with dm- to m-scale lateral extension.	Product of deposition by suspension settling. There is no evidence of possible tidal influence on deposition. Interpretation as fluid-mud deposits is excluded for occurrences in body A due to their position (cf. Dalrymple et al. 2003; Ichaso & Dalrymple 2009).

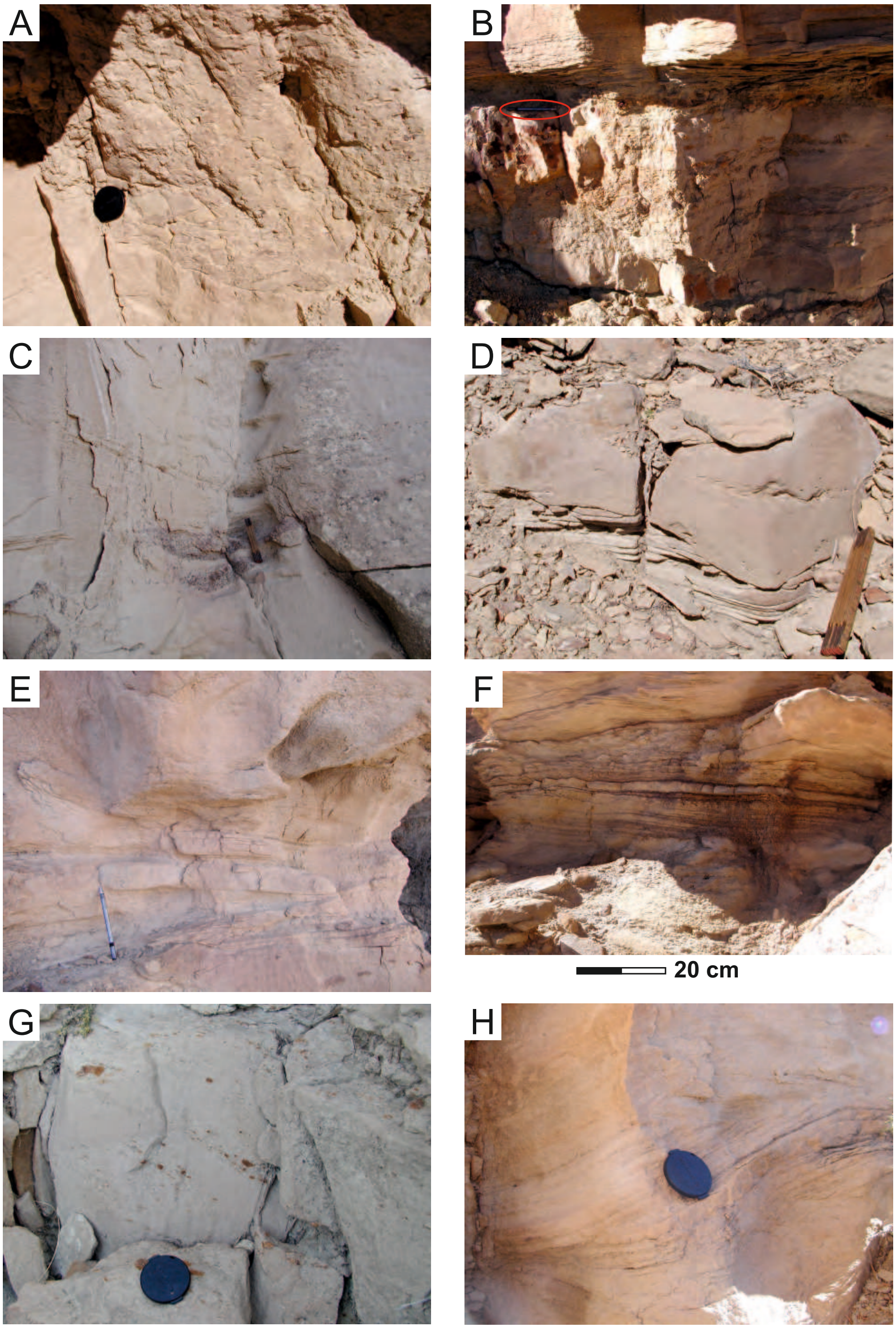
Channel-fill ID	Position of top relative to TCSB (m)	Thickness (m)	Width (m)	Lithofacies
A	21	6.5	75 (partial)	Sm, St, Sp, Sr, Sd, Fl, Fm
B	27	4.3	-	Sm, Sp
C	15	5.5	100 (real)	Gh, Gcm, Ss, Sm, Sl, Sh, St, Sp, Sc, Sr, Su, Sd
D	19	4.4	35 (real)	St, Sr, Sd
E	-3	5.4	77 (real)	Gh, Gcm, Ss, Sm, Sl, St, Sp, Sc, Sσ, Sr, Sd
F	-5	6.1	45 (partial)	Gh, Gcm, Ss, Sm, St, Sp, Sh, Sl, Sσ/c, Sr, Sd
G	17	4.8	67 (real)	Sm, Sl, St, Sr, Su, Fl

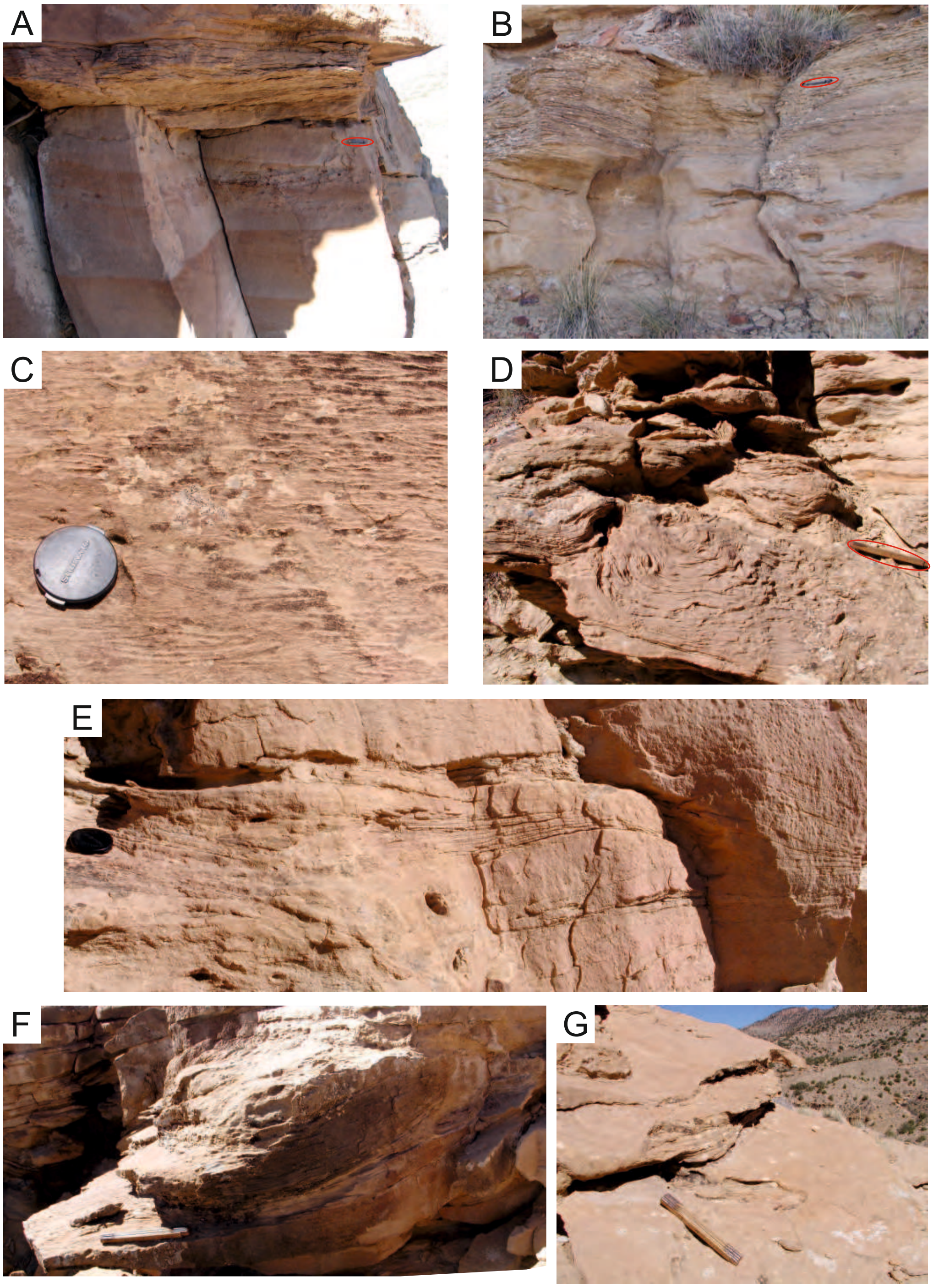

A
ALLUVIAL PLAIN
upper
DELTA PLAIN
lower
Blum et al. (2013)
MISSISSIPPI
Meander-bend migration rate (m/yr)
Channel-belt width/thickness
Distance upstream of river mouth (km)
Petter (2010)
CASTLEGATE FM.
Average paleo-flow depth (m)
Backwater zone
Basinward distance from Tusher Canyon (km)
B
ALLUVIAL PLAIN
upper
DELTA PLAIN
lower
backwater zone
drawdown zone
Flooding
Flood-related scour
ALLUVIAL PLAIN
Drawdown-related scour
LOWER DELTA PLAIN
point bar
low-stage channel
high-stage channel

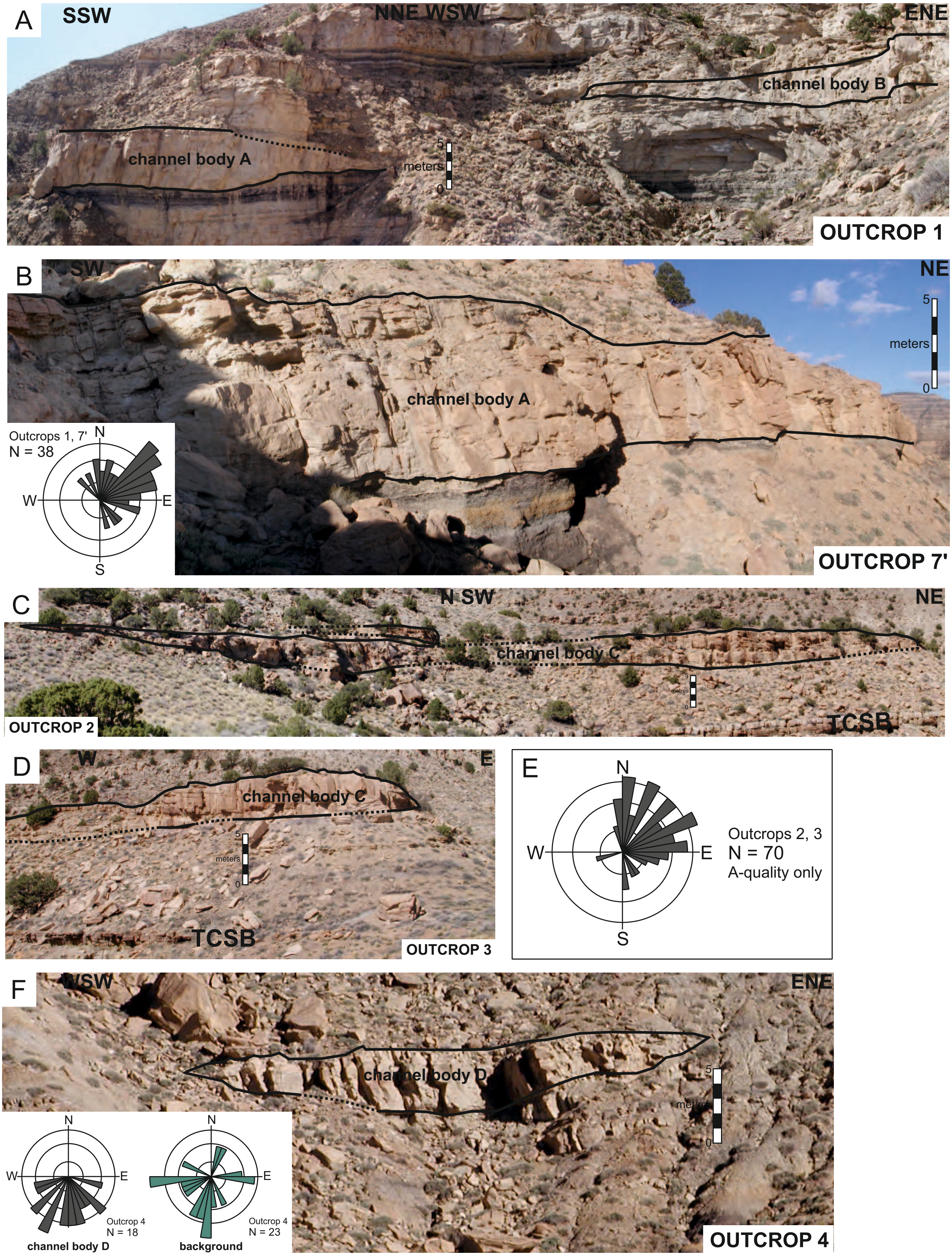

A
Cretaceous Interior Seaway
Utah
Colorado
110°
109°
Modified after Taylor et al. (2000)
Utah
Colorado
Crescent Canyon (fig. 2b)
Westwater Canyon
Book Cliffs
Green River
Tusher Canyon
To Grand Junction
39°
Green River
Sagers Canyon area
I-70
I-70
N
10 km
B
0
1
2
kilometres
2'30''
4322
4321
4320
4319
4318000mN
39°00'
602000mE
603
604
605
606
109°47'30''
1
2
3
4
5
6
7
8
Crescent Butte
Crescent Canyon
Middle Crescent Canyon


Stage
Substage
Utah
Colorado
Price Canyon
Green River
Western Book Cliffs
Eastern Book Cliffs
Grand Hogback, Coal Basin area
Maastrichtian
Early Maast.
70.6 Ma
Campanian
Late Campanian
Middle Campanian
North Horn Formation
Tuscher Formation
Ohio Creek equivalent
Undifferentiated Mbr.
Paonia Sh. Mbr.
Bowie Sst. Mbr.
Williams Fork Formation
Williams Fork Formation
Iles Fm.
Price River Formation
Farrer Formation
Rollins Sst. Mbr.
Bluecastle Tongue
Cozzette Mbr.
Study interval
Neslen Formation
Iles Formation
Corcoran Mbr.
Mesaverde Group
Castlegate Sandstone
Upper Sego Sst.
Sego Sandstone
Anchor Mine Tongue (Mancos Shale)
Lower Sego Sst.
Mancos Shale
Buck Tongue (Mancos Shale)
Castlegate Sandstone

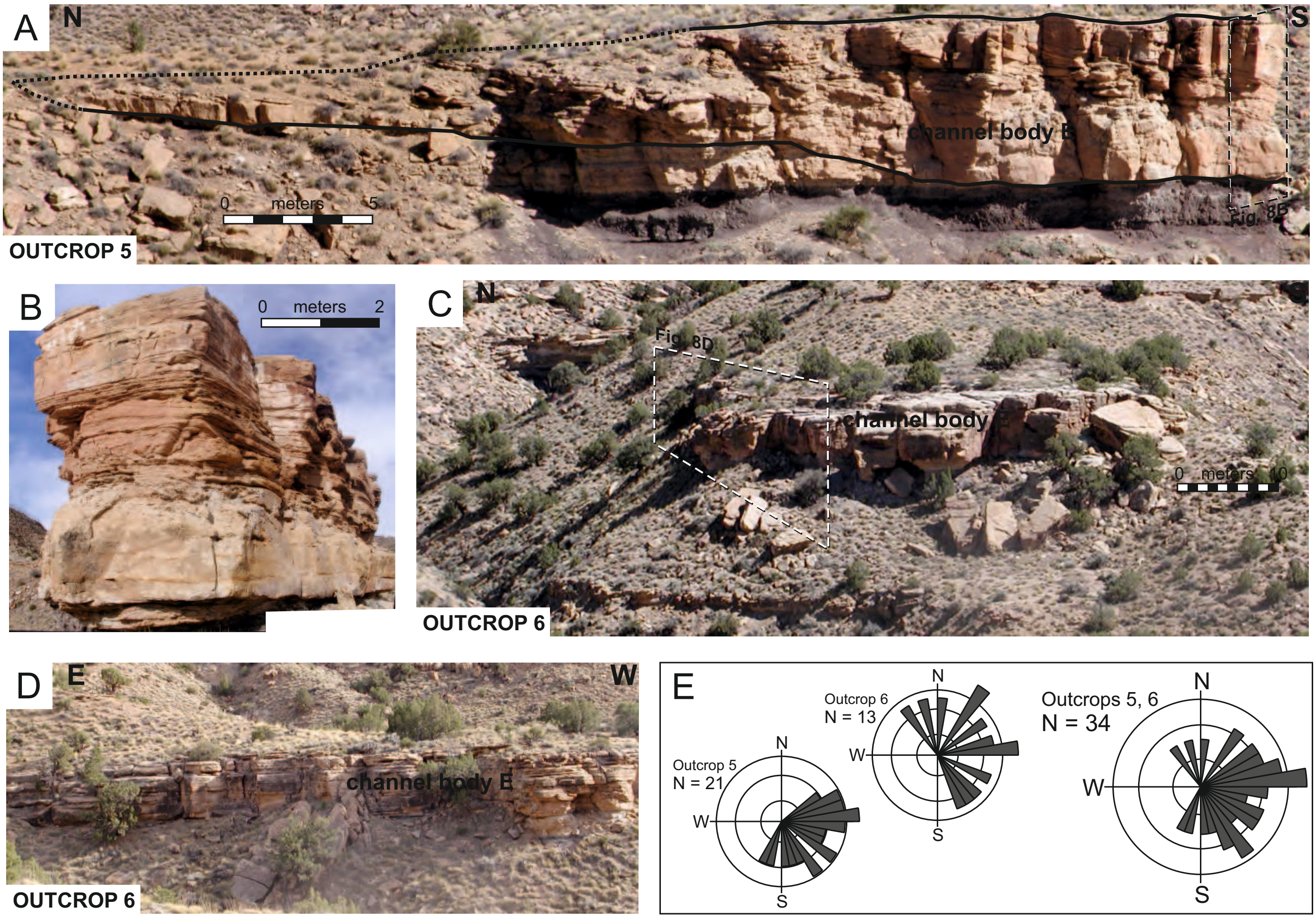

A
TCSB
LA
0 meters 5
B
LA
LA
Outcrop 6
0 meters 5
C
TCSB
CH?
LA
meters 5
D
0 centimeters 5
E
LA
N = 85
N
N = 21
N
accretion surfaces
paleocurrents
0 meters 5


A
B
C
D
E
F
20 cm
G
H

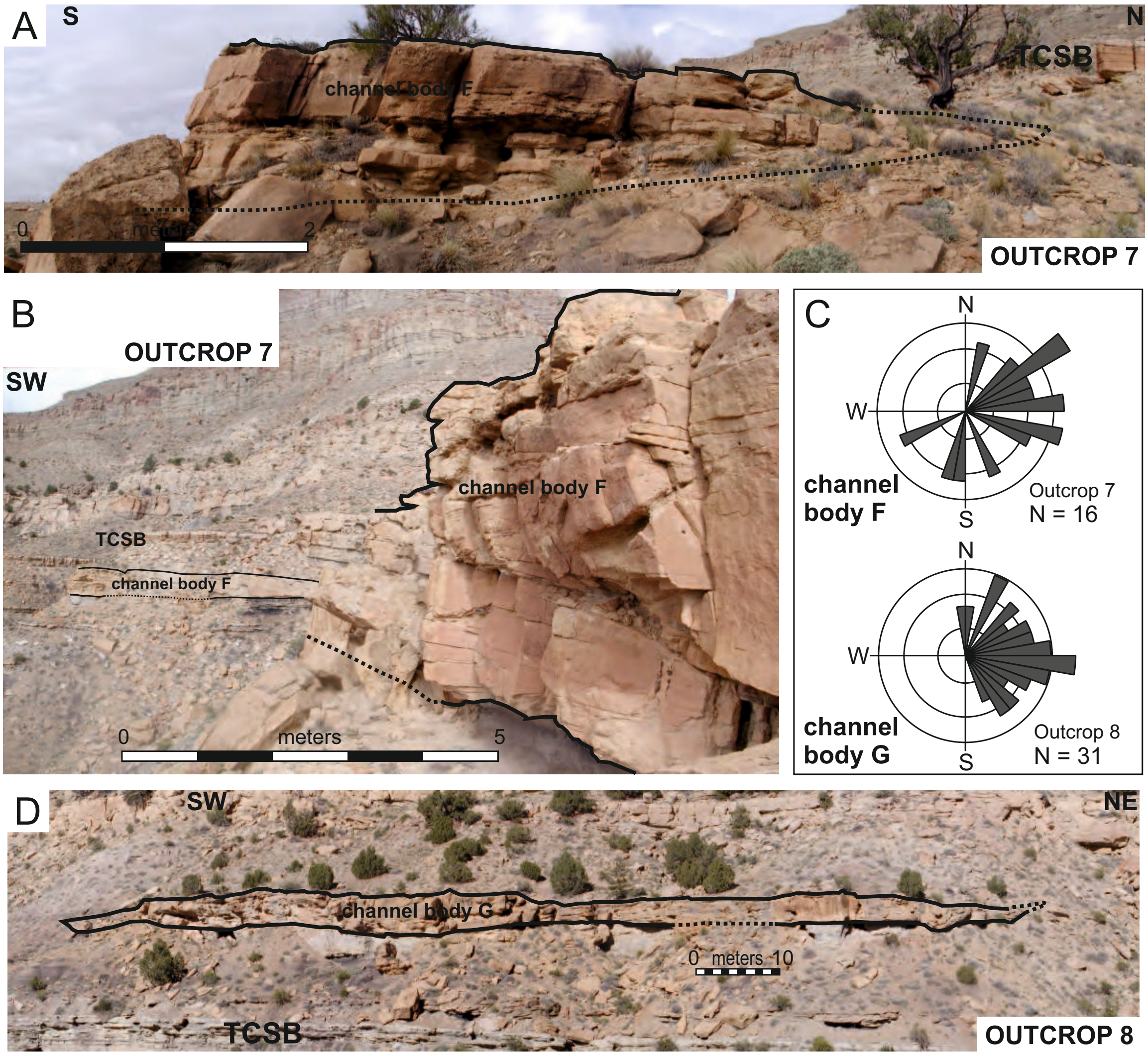

A
B
C
D
E
F
G

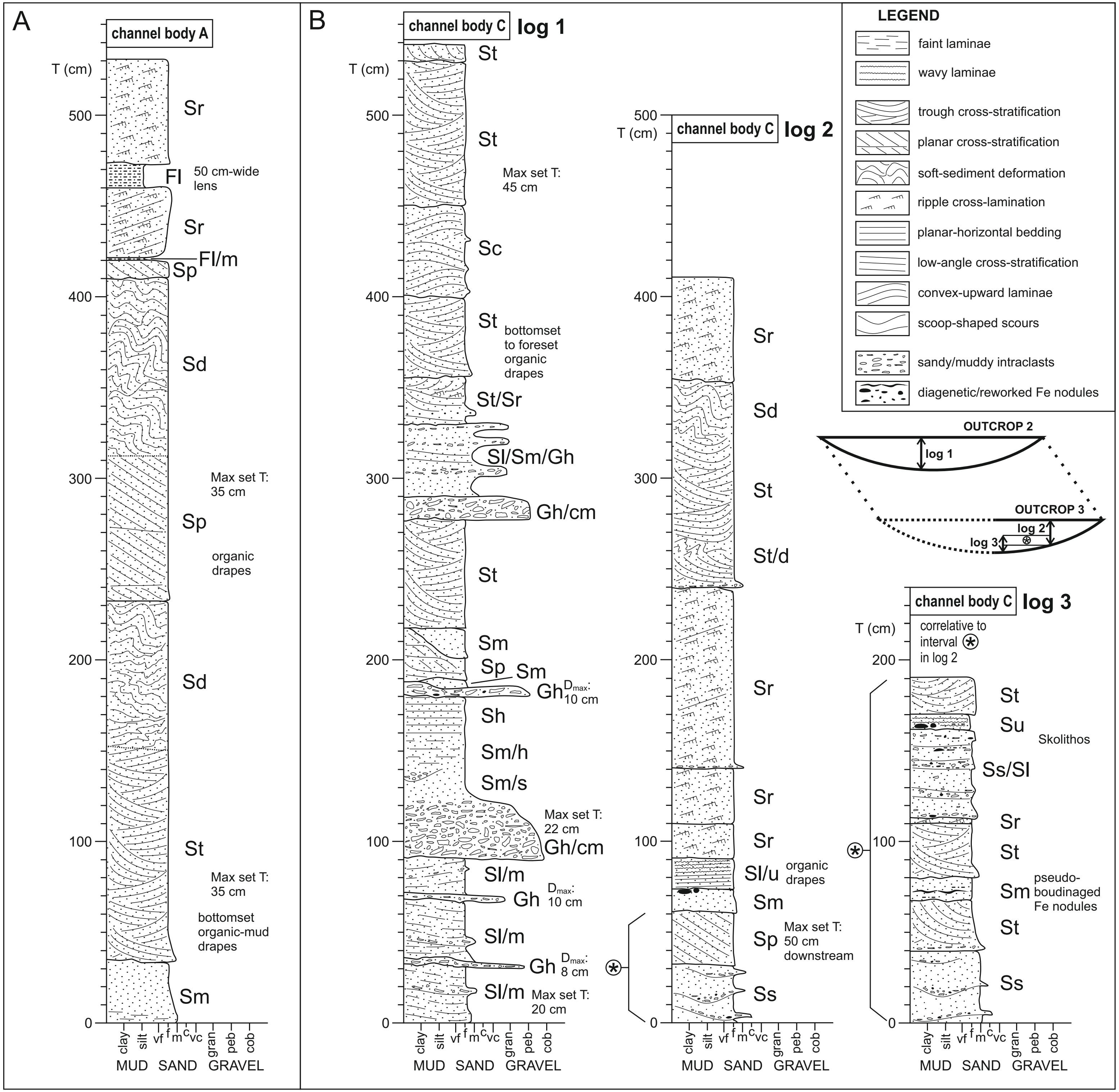


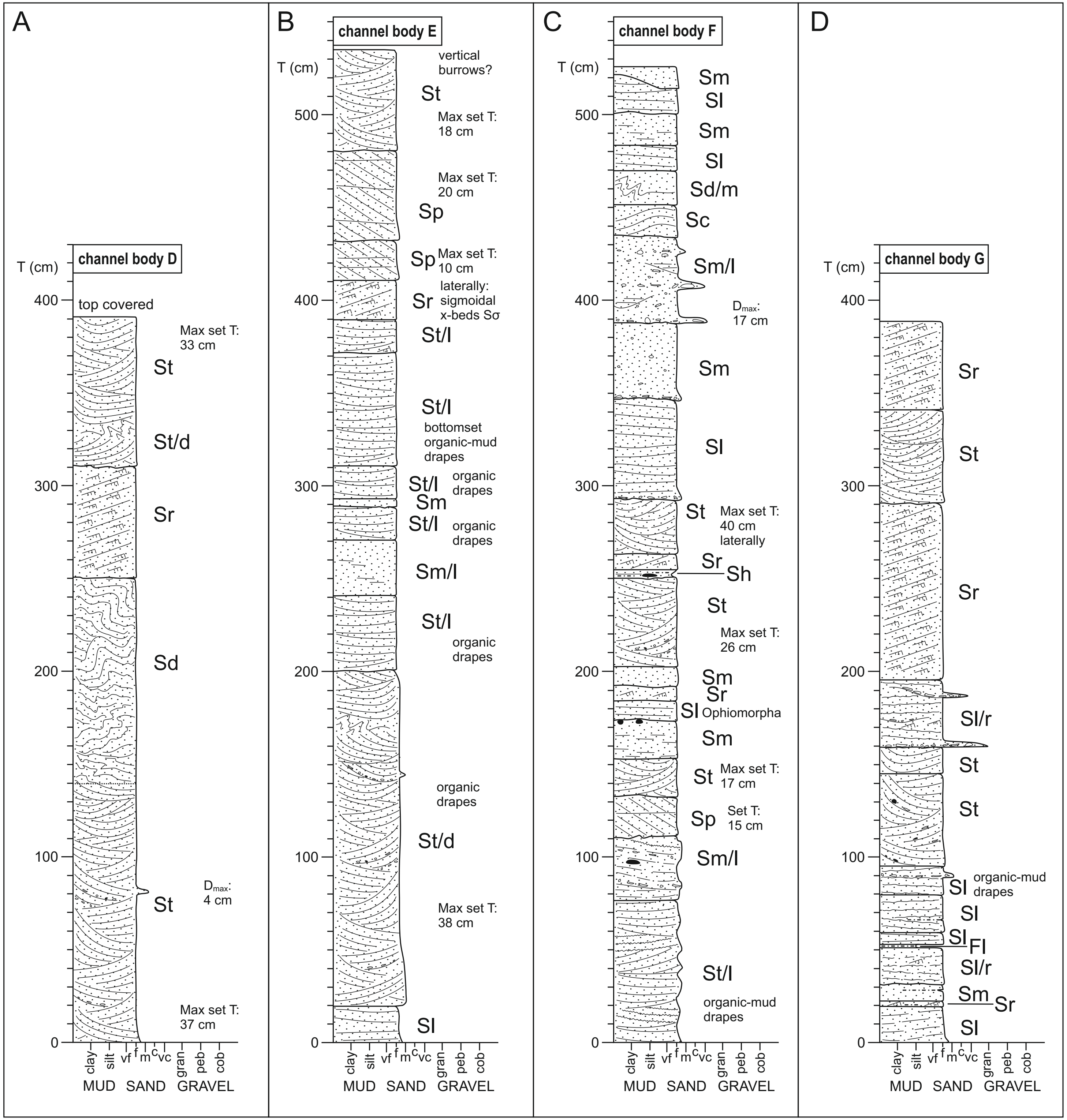
A
SSW
NNE WSW
ENE
channel body B
channel body A
5
meters
0
OUTCROP 1
B
SW
NE
5
meters
0
channel body A
Outcrops 1, 7'
N = 38
N
W
E
S
OUTCROP 7'
C
N SW
NE
channel body C
OUTCROP 2
TCSB
D
W
E
channel body C
5
meters
0
TCSB
OUTCROP 3
E
N
W
E
S
Outcrops 2, 3
N = 70
A-quality only
F
WSW
ENE
channel body D
5
0
N
W
E
Outcrop 4
N = 18
channel body D
N
E
Outcrop 4
N = 23
background
OUTCROP 4



A
N
S
channel body E
0 meters 5
Fig. 8B
OUTCROP 5
B
0 meters 2
C
N
Fig. 8D
channel body E
0 meters 10
OUTCROP 6
D
E
W
channel body E
OUTCROP 6
E
Outcrop 5
N = 21
Outcrop 6
N = 13
Outcrops 5, 6
N = 34
N
W
S


A
S
N
channel body F
TCSB
0
meters
2
OUTCROP 7
B
OUTCROP 7
SW
channel body F
TCSB
channel body F
0
meters
5
C
N
W
S
channel body F
Outcrop 7
N = 16
N
W
S
channel body G
Outcrop 8
N = 31
D
SW
NE
channel body G
0 meters 10
TCSB
OUTCROP 8


A
channel body A
T (cm)
Sr
Fl
50 cm-wide lens
Sr
Fl/m
Sp
Sd
Max set T: 35 cm
Sp
organic drapes
Sd
St
Max set T: 35 cm
bottomset organic-mud drapes
Sm
B
channel body C log 1
St
St
Max set T: 45 cm
Sc
St
bottomset to foreset organic drapes
St/Sr
Sl/Sm/Gh
Gh/cm
St
Sm
Sp
Sm
Gh
Dmax: 10 cm
Sh
Sm/h
Sm/s
Max set T: 22 cm
Gh/cm
Sl/m
Gh
Dmax: 10 cm
Sl/m
Gh
Dmax: 8 cm
Sl/m
Max set T: 20 cm
channel body C log 2
Sr
Sd
St
St/d
Sr
Sr
Sr
Sl/u
organic drapes
Sm
Sp
Max set T: 50 cm downstream
Ss
LEGEND
faint laminae
wavy laminae
trough cross-stratification
planar cross-stratification
soft-sediment deformation
ripple cross-lamination
planar-horizontal bedding
low-angle cross-stratification
convex-upward laminae
scoop-shaped scours
sandy/muddy intraclasts
diagenetic/reworked Fe nodules
OUTCROP 2
log 1
OUTCROP 3
log 2
log 3
channel body C log 3
correlative to interval in log 2
St
Su
Skolithos
Ss/Sl
Sr
St
Sm
pseudo-boudinaged Fe nodules
St
Ss
clay silt vf f m c vc gran peb cob
MUD SAND GRAVEL


A
channel body D
T (cm)
top covered
Max set T: 33 cm
St
St/d
Sr
Sd
Dmax: 4 cm
St
Max set T: 37 cm
B
channel body E
vertical burrows?
St
Max set T: 18 cm
Max set T: 20 cm
Sp
Sp
Max set T: 10 cm
Sr
laterally: sigmoidal x-beds Sσ
St/l
St/l
bottomset organic-mud drapes
St/l
organic drapes
Sm
St/l
organic drapes
Sm/l
St/l
organic drapes
organic drapes
St/d
Max set T: 38 cm
Sl
C
channel body F
Sm
Sl
Sm
Sl
Sd/m
Sc
Sm/l
Dmax: 17 cm
Sm
Sl
St
Max set T: 40 cm laterally
Sr
Sh
St
Max set T: 26 cm
Sm
Sr
Sl
Ophiomorpha
Sm
St
Max set T: 17 cm
Sp
Set T: 15 cm
Sm/l
St/l
organic-mud drapes
D
channel body G
Sr
St
Sr
Sl/r
St
St
Sl
organic-mud drapes
Sl
Sl
Fl
Sl/r
Sm
Sr
Sl
0
100
200
300
400
500
clay
silt
vf
f
m
c
vc
gran
peb
cob
MUD
SAND
GRAVEL