SEQUENCE STRATIGRAPHIC ANALYSIS OF THE YOUNGEST NONMARINE SEQUENCE IN THE CRETACEOUS FERRON NOTOM DELTA, SOUTH CENTRAL UTAH, U.S.A.

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ABSTRACT: This study evaluates the nature and internal complexity of high-frequency fluvial aggradational cycles and linked paleosols that were previously not recognized in nonmarine sequences of the Turonian Ferron Notom Delta in Central Utah. Detailed facies analysis of the ~ 33-meter-thick youngest nonmarine sequence (Sequence 1) reveals 11 depositional facies that build 33 fluvial aggradational cycles, 9 fluvial aggradation cycles sets, and 3 sequences, in what was previously classified as a single 100 ka, 4th- or 5th-order sequence. Fluvial aggradation cycles in the succession are either simple floodplain bedsets or single channel stories that are bounded by paleosols or erosional surfaces. They represent depositional events that likely do not span more than 2 ka. Depositional periods are followed by short hiatus (10 to < 400 a) characterized by subaerial exposure, and soil development, possibly reflecting floodplain abandonment or short-term river avulsions. Fluvial aggradational cycle sets show systematic stacking of fluvial aggradational cycles, and are analogous to parasequence sets. They are represented by either compound floodplain bedsets or multi-story channel sandstones, bounded by relatively mature paleosols or erosional top-truncated. They represent multiple small-scale depositional events that span not more than 14 ka followed by relatively long (100 a to < 1 ka) hiatus, characterized by subaerial exposure and soil development, which suggest a relationship to longer-term floodplain abandonment, due to either longer-term avulsion or fluvial entrenchment during base-level falls. Fluvial aggradational cycle sets are often capped by coals and carbonaceous strata. Marine trace fossils and dinocysts above coal beds suggest marine transgression due to compactional subsidence or sea-level rise. The scale of cyclicity of the fluvial aggradation cycle sets (14 ka) is similar to underlying marine parasequences (10 to 15 ka) and suggests that nonmarine and marine cycles have a similar control. Fluvial sequences are aggregates of fluvial aggradational cycle sets that are bounded by unconformities or their correlative conformities that represent longer periods (1 to 5 ka) of subaerial exposure and pedogenic development. Cyclicity on the sequence scale in this succession is attributed to high-frequency Milankovitch-scale (~ 20 ka, 40 ka) rises and falls in sea level.

INTRODUCTION

There have been several attempts to identify cyclic hierarchies in fluvial strata (e.g., Allen 1974; Bridge 1984; Kraus 1987; Willis and Behrensmeyer 1994; Kraus and Aslan 1999) that broadly match the hierarchical division of marine strata into parasequences, parasequence sets, and sequences (Van Wagoner et al. 1990; Shanley and McCabe 1994; Wright and Marriott 1993). For example, Atchley et al. (2004) integrated both channel and floodplain deposits in their study of Upper Cretaceous to lower Tertiary nonmarine strata in west Texas and identified three hierarchies of alluvial cyclicity in which fluvial sequences comprise fluvial aggradational cycle sets (FAC sets) built of the smallest-scale fluvial aggradational cycles (FACs). Prochnow et al. (2006) and Cleveland et al. (2007) respectively, applied this method in characterizing the Upper Triassic Chinle Formation in Utah and New Mexico, but there have been few similar studies. These previous studies did not have associated or equivalent marine strata, and thus it was not possible to compare the cyclicity of alluvial and marine depositional sequences. Also, these previous studies did not have high-resolution chronometric controls in the respective studied intervals, making time-stratigraphic analysis challenging and precluding a fuller explanation of allogenic and autogenic mechanisms for the generation of the facies cycles.

This study examines the youngest sequence in the Turonian-age Ferron Notom Delta complex: a nonmarine sequence that has been previously subdivided into three nonmarine “parasequences” based on broad channel stacking patterns and identification of possible brackish facies linked to coastal-plain flooding events (Zhu et al. 2012). The Zhu et al. (2012) study was regional in nature and did not undertake detailed analysis of the cyclicity in the associated channel and floodplain deposits, but it did incorporate bentonite chronometry that provided a high-resolution time-stratigraphic framework.

The purpose of this study is to conduct a more detailed facies and stratigraphic analysis of this nonmarine sequence by: (1) incorporating detailed descriptions of channel and floodplain facies, (2) identification and analysis of fluvial aggradational cycles and associated paleosols, using the approach of Atchley et al. (2004), (3) correlation of cyclic depositional events across a several kilometer-long stretch of outcrop, (4) comparison of this cyclicity to the hierarchy of marine cycles within the Ferron Notom Delta complex, and (5) evaluation of allogenic and autogenic controls on development of high-frequency (sub-Milankovitch) nonmarine cycles.
The study also attempts to highlight the stratigraphic complexity in characterizing fluvial strata and the recognition of stratal hierarchies that may have been previously overlooked. It highlights the importance of integrating flood-plain deposits in the stratigraphic analysis of fluvial strata, which often involves the usual practice of observing or interpreting strata based on channel stacking patterns.

GEOLOGICAL SETTING AND PREVIOUS WORK

The Western Interior Foreland Basin developed between the North American Craton to the east and the Cordilleran volcanic arc to the west as a flexural response to eastward-directed thrust-sheet loading of the Sevier orogenic belt (DeCelles et al. 1995; DeCelles and Giles 1996; Ryer and Anderson 2004; Edwards et al. 2005). Rapid subsidence of this foredeep on the eastern side of the Sevier orogenic belt spanned most of the Cretaceous period (DeCelles and Giles 1996; Ryer and Anderson 2004). The Western Interior Seaway filled the foredeep as a result of continued subsidence and eustatic sea-level rise during the Cretaceous.

This epeiric seaway extended from the Boreal Sea in the north, to the Gulf of Mexico (Uresk 1978). A series of clastic wedges, along the margin of the seaway, sourced from the adjacent Sevier orogenic belt, prograded eastward into the seaway (Bhattacharya and MacEachern 2009; Fielding 2010, 2011) (Fig. 1). The Turonian Ferron Sandstone Member of the Mancos Shale Formation comprises three clastic wedges (i.e., the Last Chance, the Vernal, and the Notom) (Cotter 1976; Uresk 1978; Hill 1982; Bhattacharya and Tye 2004; Ryer and Anderson 2004; Zhu et al. 2012). Paleogeographic reconstruction shows a 40–50°N paleolatitude for the Ferron clastic wedges during the Turonian (Bhattacharya and MacEachern 2009), suggesting a subtropical climate due to expansion of tropical climate conditions into higher latitudes during the Cretaceous greenhouse climate.

The Ferron Notom clastic wedge crops out in the Henry Mountains and Caineville areas of south-central Utah. It is bounded below by the Tununk Shale and unconformably above by the Santonian Blue Gate Shale, both members of the Mancos Shale Formation (Fig. 2; Peterson and Ryder 1975). The Ferron Notom wedge was informally divided into an upper fluvial and a lower marine unit (Peterson and Ryder 1975; Uresk 1978; Hill 1982). The lower unit consists of shoreface and delta-front sandstones and marine shales, whereas the upper unit consists of channel sandstones, flood-plain deposits, and coals (Peterson and Ryder 1975; Hill 1982; Fielding 2010, 2011; Li et al. 2010). However, Zhu et al. (2012) showed that the upper fluvial unit contains an interfingering transgressive marine tongue (Fig. 3A), and recently Richards (2014) showed additional intertonguing of the marine and fluvial units. This study focuses on the uppermost fluvial unit above predominantly marine facies. Regional work shows that the largest trunk streams in the Ferron Notom wedge were ~7 m deep. Smaller streams are interpreted as either distributary channels, if unconfined (e.g., Ahmed et al. 2014; Li and Bhattacharya 2013; Bhattacharyya et al. 2015), or as tributary systems associated with the flanks of incised valleys (Li et al. 2010; Hilton 2013; Li and Bhattacharya 2014).
Li et al. (2010) and Zhu et al. (2012) conducted a regional sequence stratigraphic study of the Ferron Notom Deltaic Complex (Fig. 3A, B). Their regional cross sections reveal 1 sequence set, 5 sequences, 18 parasequences, and 43 parasequences. Based on $^{40}$Ar/$^{39}$Ar isotopic dating of sanidine crystals from 4 bentonite horizons within and bounding the sedimentary package, including a bentonite associated with coal zone 2 in Sequence 1, Zhu et al. (2012) determined the duration of deposition of the entire Ferron Notom wedge to be 620,000 years. Age dates from the middle of Sequence 1 (coal zone 2) to the base of sequence 2 (Parasequence 7) indicate a duration of about 50,000 years (Zhu et al. 2012). Using simple averaging, the duration of each sequence is assumed to be on the order of $10^5$ years, indicating that they represent high-frequency (4th- or 5th-order) Milankovitch-scale sequences.

The upper two sequences in the Ferron Notom wedge (sequence 1 and sequence 2) are marked by compound valley systems (Fig. 3A, B). The uppermost incised valley constitutes the lower part of the youngest sequence 1. This compound valley truncates marine strata in more distal positions (Fig. 3A) and fluvial deposits of the underlying sequence 2 in more proximal areas (Zhu et al. 2012). This uppermost valley is marked by three separate cut-and-fill episodes, each of which shows evidence of up to 30 m of erosion and filling (Li et al. 2010; Li and Bhattacharya 2013). These higher-frequency cut-and-fill episodes are interpreted as higher-frequency Milankovitch cycles (e.g., 20 ka to 40 ka) with the smallest-scale successions probably reflecting autogenic processes, such as floodplain abandonment, short- and long-term river avulsion, or compactional compensation cycles. This is explored in the discussion and interpretation sections of the paper.

**STUDY AREA AND METHODOLOGY**

The study area is in the Sweetwater Creek drainage, between the Henry Mountains and Utah Highway 24 (Figs. 4, 5). This location was chosen for its continuous cliff exposures of well-preserved channel sandstone and floodplain deposits. The northern part of the study area exposes all of the youngest Ferron Notom Sequence 1 from the base, where it truncates underlying marine strata (Parasequence 4 of Li et al. 2010, Zhu et al. 2012, and Li and Bhattacharya 2013) to the Blue Gate Shale, which caps the succession (Fig. 6A). The southern area exposes the upper part of the sequence (Fig. 6B). The southern and northern areas cover a continuously exposed distance of about 1800 m and 500 m, respectively (Fig. 5). The distance between the two study areas is about 2 km, with variable exposure in between. The thickness of the entire succession is ~33 m.

A total of 29 detailed stratigraphic sections were measured, 25 of which were measured in South Sweetwater Creek drainage and 4 were measured in the North Sweetwater Creek drainage. Locations of measured sections were marked by GPS, and were spaced according to distances indicated by GPS waypoints. All measured sections (e.g., Figs. 7, 8) contain detailed outcrop descriptions, which include lithofacies, grain size, stratal thickness, stratigraphic boundaries or discontinuities, trace fossils, and organic-matter content.

Mudstones and paleosols were described based on their color, structure, and other constituents (e.g., root traces, slickensides, nodules, mottles, and organic-matter content). The intensity of rooting or rooting index of paleosols was also tracked and recorded. Determination of rooting index (R.I.) of a paleosol in the succession was based on visual comparison of strata without any root traces (R.I. = 0) to well-rooted paleosol horizons (R.I. = 5), and a rooting index of 2 or 3 for moderately rooted paleosol horizons.

A total of 356 paleocurrent measurements were obtained from cross-bed foresets, plan-view rib-and-furrow structures, and ripple cross lamina in channel-fill sandstone facies. Paleo-discharge calculations followed methods of Bridge and Tye (2000) and Bhattacharya and Tye (2004). Thicknesses of preserved channel and channel-bar deposits were obtained to estimate bankfull depth of paleo-channels. The thickness of channel bar deposits is roughly 80-90% of channel depth (Bridge and Mackey 1993; Bridge 2003).

Apparent widths of channel sandstones were obtained from outcrop data and were adjusted to approximate true widths, consistent with outcrop orientation and paleocurrent data. The approximate widths were compared with results from empirical width-estimation methods, including:
FIG. 3.—Regional sequence stratigraphy of the Ferron Notom Delta in cross section. A) Depositional-dip cross section (Zhu et al. 2012). B) Depositional-strike section (Li et al. 2010). See Figure 4 for both cross-section outlines. This study focused primarily on Sequence 1, which is the uppermost and youngest sequence (interval indicated by the red boxes).
FIG. 4.—Maps showing the Ferron Notom outcrop in south-central Utah. Red lines represent regional dip (X–X’) and strike (Y–Y’) cross sections in Figure 3. Blue box is the study area, and black dots represent the cross section outline, which are expanded in Figure 21 (modified after Zhu et al. 2012).
Bridge and Mackey (1993a):

\[ w_c = 8.8(d_m)^{1.82} \]  
(1)

\[ w_c = 59.86(d_m)^{1.8} \]  
(2)

and Fielding and Crane (1987):

\[ w = 64.6(d_m)^{1.54} \]  
(3)

where \( w_c \) and \( w \) = channel width, and \( d_m \) = mean channel depth. In cases where widths of channel sandstones could not be obtained, due to outcrop orientation limitations, these numerical methods were applied to estimate the approximate channel widths. Channel cross sectional area \( (A) \) was then calculated using

\[ A = 0.65(w \times d) \]  
(4)

where \( d \) = channel depth, and multiplication of the cross sectional area by 0.65 accounts for the curvature of channels, as they are not fully rectangular (Bhattacharya and MacEachern 2009). Flow velocity \( (U) \) was estimated using the three-dimensional bedform phase diagram of Rubin and McCulloch (1980) based on grain size, bankfull channel depth, and dominant sedimentary
structure in respective channel sandstones. Bankfull paleo-discharge ($Q_w$) was then calculated using

$$Q_w = A \cdot U$$  \hspace{1cm} (5)

Correlation of measured sections was carried out by physically walking and tracing bed-to-bed contacts and lateral extents of individual bedding surfaces or channel belts in order to produce detailed stratigraphic cross sections of the two areas. The stratigraphic cross sections are both oriented slightly oblique to depositional strike.

**LITHOFACIES DESCRIPTION**

A total of 11 lithofacies were identified and described and are summarized in Table 1 and Table 2, where they have also been divided into those that consist predominantly of sandstones and those that contain predominantly mudstones and siltstones (Figs. 9–15). From here onwards facies will be often be referred to by the number that each has been allotted in the respective tables.

However, it is important to discuss the importance of paleosol (FA. 9; Fig. 15) and coal (FA. 11; Fig 15A) identification to this study because they served as useful tools for the identification of stratal packages in the studied interval. These are discussed below:

**Paleosol Maturity**

The general criteria for identifying paleosols are horizonation, ped (soil aggregates) structure development, rooting, disruption of original sedimentary fabric, and accumulation of pedogenic clay (Retallack 2001). Paleosols identified in this study based on these criteria show variation in the development of each of the above-mentioned features (Fig. 15), and this was used for the determination of their level of maturity. Based on these development criteria, five types of paleosols were identified and are listed in Table 3. Excluding Type I, which are predominantly coals, they are listed based on their level of development and maturity. These paleosols are equivalent to modern Entisols, Inceptisols, Vertisols, and Alfisols (Soil Survey Staff 1999).

Generally, paleosols in this study show distinctive low-chroma gray colors with light-gray, pale-orange, or pale brownish-yellow mottles (Fig. 15A), which are attributed to gleying; a product of iron or manganese reduction by anaerobic bacteria under poorly drained, oxygen-limited conditions (Retallack 1997, 2001; McCarthy et al. 1999; Driese and Ober 2005). Clay-coated surfaces of paleosols often have a waxy luster and may have striations and

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**Fig. 6.—The outcrop of Sequence 1 at Sweetwater Creek. A) The full sequence is exposed at North Sweetwater from the underlying marine parasequence to the top, and it is capped by the Blue Gate Shale. B) In South Sweetwater the underlying marine parasequence and the lower parts of the sequence are not exposed. Note the person (indicated by red arrow) for scale.**
grooves (Fig. 15C). These are pedogenic slickensides that form due to repeated shrinking and swelling caused by alternating wet (poor drainage) and dry (better drainage) phases associated with seasonal water-table fluctuations (Brewer 1976; Leckie et al. 1989; Retallack 1997, 2001; McCarthy et al. 1999; Driese and Ober 2005). They are common in mudstone paleosols.

Laterally extensive coal beds underlain or overlain by carbonaceous mudstones or siltstones were also identified in the succession, and they commonly overlie floodplain paleosols. Thickness of coal beds vary from 5 to 40 cm, and they extend for hundreds of meters to several kilometers beyond the study area. Coals and associated carbonaceous mudstones or siltstones, referred to as coal zones in this study, are analogous to modern Histosols (Table 3; Type I; Soil Survey Staff 1999). The rank of coals observed in this study is lignite to sub-bituminous, suggesting an early or immature stage of coal development. The trace fossils Teredolites and Thalassinoides were found in abundance above and below some coal beds (Fig. 12C, D), suggesting the formation of a marine-influenced woodground (Bromley et al. 1984; Savrda 1991; MacEachern and Pemberton 1992; Benton and Harper 1997; Bohacs and Suter 1997; Gingras et al. 2004). Coal zones were useful correlation tools in this study, and their significance will be more evident later in this discussion.

**Cyclic Hierarchy and Sequence Stratigraphy**

In this study, small- to large-scale genetically related fluvial stratigraphic successions bounded by paleosols or erosional surfaces have been identified, based on the vertical and lateral relationships of the facies, described above. Similar types of fluvial stratigraphic successions have been previously described by Bridge (1984), Kraus (1987), Willis and Behrensmeyer (1994), and Aslan and Autin (1999; their fig. 8). Atchley et al. (2004, 2013), Prochnow et al. (2006), and Cleveland et al. (2007), in hierarchical order, described these as: fluvial aggradation cycles (FACs), fluvial aggradation cycle sets (FAC sets), and fluvial sequences, and this terminology is also adopted in this study. Table 4 provides a summary of similar observed stratigraphic hierarchies identified in the studied interval. The identification criteria, characteristics, and interpretation of hierarchical units are presented below.

**Fluvial Aggradation Cycles**

Fluvial aggradation cycles (henceforth FACs) observed in the Ferron Notom are mappable decimeter- to meter-scale units that are bounded either by pedogenically modified surfaces of rooting and burrowing or by surfaces representing erosional truncation. A single FAC is represented by a paleosol-capped or erosionally top-truncated decimeter- to meter-thick single-storied channel sandstone, or a decimeter- to meter-thick simple floodplain bedset. An example is shown in Figure 16A, where a coarsening-upward floodplain bedset, consisting of a basal shallow floodplain-lake fill (FA. 8), and an upper fine-grained crevasse-splay deposit (FA. 4), is capped by a rooted paleosol (FA. 9). Figure 16C shows another example of an underlying and truncated stacked channel bedset (FA. 2) representing a single FAC. The upper stacked channel bedset capped by a paleosol represents another FAC. A total of 33 distinct FACs were identified in the studied interval, and their thickness varies from about 0.5 m to 6.5 m for the most representative units.
FIG. 8.—Type measured section of representative fluvial facies successions in Sequence 1. The contact between the basal marine shoreface facies (FA. 1) and the overlying channel-fill sandstone facies marks the base of Sequence 1. The top of the sequence is at the marine Blue Gate Shale contact. See legend in Figure 7.
Table 1.—Summary of predominantly sandstone facies recognized in the studied interval.

<table>
<thead>
<tr>
<th>Facies (FA.)</th>
<th>Lithology</th>
<th>Sedimentary Structures and Depositional Profile</th>
<th>Biota</th>
<th>Geometry and Architectural Element</th>
<th>Depositional Feature</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Very fine- to fine-grained bioturbated sandstone</td>
<td>Hummocky cross stratification, wave-ripple lamination, and planar stratification. Moderate to high bioturbation.</td>
<td>Opalina, Paleophycus, Thalasinosoides, Planolites, Ruelia, and Cylindrichnus.</td>
<td>Mainly tabular and extensive sandstones. Erosionally truncated top (Figs. 8, 20).</td>
<td>Middle to lower Shoreface *Ref.: Galloway and Hobday 1996; Clifton 2006; Fielding 2010</td>
</tr>
<tr>
<td>2</td>
<td>Very fine- to very coarse-grained sheet or lenticular sandstone</td>
<td>Large-scale inclined stratification, small- to large-scale trough cross-bedding, planar stratification, and current-ripple cross-lamination. Convolute bedding. Single or multiple (often truncated) upward-finishing profiles. Basal coarse-grained or pebble lags and intraformational mud clasts.</td>
<td>*Ref.: Stands for References.</td>
<td>Erosionally based, extensive, amalgamated sheet sandstones with dipping unit and compound bars. Lateral lenticular sandstones. Some show dipping cm- to dm-thick accretion sets (Figs. 8, 10).</td>
<td>Channel Sandstone *Ref.: Bristow and Best 1993; Miall 1996; Aslan and Autin 1999; Bridge 2006; Bridge and Lant 2009</td>
</tr>
<tr>
<td>3</td>
<td>Very fine- to fine-grained sandstone and interbedded mudstone (or heterolithics)</td>
<td>Small- to medium-scale trough cross-bedding, current-ripple cross-lamination, lenticular, wavy, and chaotic bedding, mud drapes (with some paired). Low to moderate bioturbation. Upward-finishing profile.</td>
<td>Opalina, Pleurodictyum, and Thalasinosoides.</td>
<td>Erosionally-based gently to steeply dipping accretion sets that consists of cm- to dm-thick sandstones and interbedded mudstones (Fig. 8, 11).</td>
<td>Tide-influenced channel sandstone *Ref.: Reineck and Wunderlich 1968, deRaaf and Boersma 1971; Terwindt 1971; Visser 1980; Savrda 1991; Benton and Harper 1997; MacEachern and Pemberton 1992; Martin 2000; Gingras et al. 2004</td>
</tr>
<tr>
<td>4</td>
<td>Very fine- to medium-grained sheet sandstone</td>
<td>Small- to medium-scale cross-bedding, current-ripple cross-lamination. Mostly upward-coarsening, some upward-finishing, and few indefinite profiles.</td>
<td>Root traces at the uppermost portion or may penetrate the entire thickness.</td>
<td>Gradational to abrupt base. Sheet or lobate sandstones that consist of cm- to dm-thick sandstones and cm-thick mudstones (or rhythms) (Figs. 8, 13A, B).</td>
<td>Crevasse-splay channel and lacustrine-delta fill *Ref.: Allen 1965; Bridge 1984, 2003, 2006; Farrell 1987; Kraus 1987; Tye and Coleman 1989; Smith et al. 1989; Miall 1992, 1996; Willis and Behrensmeyer 1994; Aslan and Autin 1999; Bristow et al. 1999</td>
</tr>
<tr>
<td>5</td>
<td>Very fine- to fine-grained wedge-shaped sandstone</td>
<td>Predominantly current-ripple cross-lamination. Rare basal load casts</td>
<td>Cylindrichnus, Naktodemas, Anorichnus, and Steinichnus (Fig. 14).</td>
<td>Wedge-shaped sandstones that consist of cm- to dm-thick sandstones and interbedded mudstones (Fig. 13C, D)</td>
<td>Levee *Ref.: Fisk 1944; Allen 1965; Bridge 1984, 2003; Frey et al. 1984; Farrell 1987; Miall 1996; Brierley et al. 1997; Aslan and Autin 1999; Pérez Archea and Smith 1999; Hasiots 2002, 2004</td>
</tr>
</tbody>
</table>

FAC Interpretation

The deposition of the paleosol-capped bedset in Figure 16A can be interpreted as the result of a series of successive events, which commenced with flooding and development of a shallow ephemeral floodplain lake (FA 8), followed by levee breaching and crevasse-splay progradation, which filled the shallow lake with a lacustrine mouth bar (FA 4). Figure 19A clearly illustrates this process. A period of stability marked by cessation in flooding and deposition, and subsequent abandonment led to pedogenic development (FA 9). In the other example (Fig. 16C), a period of channel bedset (FA 2) deposition was followed by abandonment either due to local avulsion (Fig. 19B) or cut-off in a meandering channel, subsequent truncation probably occurred due to channel re-occupation by the overlying channel FAC (Fig. 19C) (Aslan et al. 2005). When there is no truncation, pedogenic development follows abandonment (FA 9). Therefore, fluvial aggradation cycles are interpreted to record episodes of floodplain aggradation and channel deposition followed by either cessation in flooding, or abandonment marked by subaerial exposure and paleosol development.

Soils that form during abandonment are often immature because of short-lived pedogenic development before the next flood event or avulsion. Paleosols that cap FACs in this study have poor to very poor horizon development. Paleosols of such FACs in this study have poor to very poor horizon development, poorly developed ped structures, little to moderate degree of rooting (R.L. = 1–2 ), existing sedimentary stratification, and few to no pedogenic clay accumulations. Variation in soil development observed is probably influenced by drainage during this phase (Aslan and Autin 1999), where soils that develop at higher elevations in the floodplain (e.g., on channels, levees, proximal
crevasse splays) are better drained and are more well developed than those in areas of lower elevation (e.g., backswamp, distal crevasse splay, floodplain lake fills, and abandoned channels), which are often submerged.

**Time Duration of FACs**

In order to compare the previous estimate of the approximate time of development (i.e., <10³ years) for FACs, sediment accumulation rates in modern rivers that drain adjacent mountain belts, analogous to the Ferron Notom, were considered (e.g., the Po River (Italy), the Ebro River (Spain), and the Rhone River (France). Bhattacharya and Tye (2004) previously suggested that these rivers represent appropriate modern analogs for the Ferron in terms of location, size of catchment area, and mean annual discharge. Sediment accretion rates reported for these rivers over periods as short as four years and as long as seventy five years vary from 4 to 30 mm/yr (Frignani and Langoone 1991; Ibanez et al. 1997; Hensel et al. 1999). This wide variation is likely a reflection of the length of the sediment accretion record, as well as the difference in sedimentation rates between areas proximal to the main river channel and the distal floodplain environment. For example, Hensel et al. (1999) differentiated sedimentation rates in areas proximal to the main Rhone River channel, which have high sedimentation rates (13.4 ± 7.0 mm/yr), from areas distal to the main channel (i.e., wetlands) with low sedimentation rates (1.1 ± 0.1 mm/yr). Similar observations were also made by Rostan et al. (1997) in the Rhone River.

In a previous study, Zhu et al. (2012) divided the total average thickness of the six sequences in the Ferron Notom (130 m) by the total time of deposition (620 kyr) and suggested a sedimentation rate of 0.2 mm/yr for the entire Ferron succession. However, this study covers only one of those sequences (sequence 1 with a thickness of 33 m) in great detail, and has incorporated the finer-scale depositional elements (i.e., floodplain deposits) into this analysis. In addition, an average sedimentation rate of 0.2 mm/yr (Zhu et al. 2012) is probably too low for a steep-gradient river system that drained an adjacent mountain belt in a humid subtropical to tropical climate with a probable high amount of precipitation and high storm influence (Milliman and Syvitsky 1992; Bhattacharya and MacEachern 2009).

Therefore, based on the sedimentation rates from modern Ferron analogs, and thickness of FACs (0.5 to 6.5 m) in the studied interval, the estimated time for channel or floodplain deposition varies from 16 years to 1,600 years. The duration of floodplain and channel abandonment and subaerial exposure can be estimated from the degree of pedogenic development on the abandoned floodplain and channel substrates. The paleosols that cap FACs are mostly immature and are similar to modern Entisols (Type II), and early-stage Inceptisols (Type III) and Vertisols (Type IV), which develop in a few tens to a few hundreds of years (U.S. Soil Survey Staff 1999; Retallack 1988, 2001; Aslan and Autin 1999; Autin and Aslan 2001; Dixon and Ober 2005). Based on these estimates, FACs in this interval likely represent no more than a few tens to a few hundreds of years (U.S. Soil Survey Staff 1999; Retallack 1988, 2001; Aslan and Autin 1999; Autin and Aslan 2001; Dixon and Ober 2005). Based on these estimates, FACs in this interval likely represent no more than a few tens to a few hundreds of years (U.S. Soil Survey Staff 1999; Retallack 1988, 2001; Aslan and Autin 1999; Autin and Aslan 2001; Dixon and Ober 2005). Based on these estimates, FACs in this interval likely represent no more than a few tens to a few hundreds of years (U.S. Soil Survey Staff 1999; Retallack 1988, 2001; Aslan and Autin 1999; Autin and Aslan 2001; Dixon and Ober 2005).

**Table 2.** Summary of predominantly mudstone facies recognized in the study interval.

<table>
<thead>
<tr>
<th>Facies (FA)</th>
<th>Lithology</th>
<th>Color</th>
<th>Sedimentary Structures</th>
<th>Biota</th>
<th>Depositional Feature</th>
</tr>
</thead>
<tbody>
<tr>
<td>6</td>
<td>Organic-rich, mudstone, silstone, and very fine-grained sandstone</td>
<td>Light to dark brown.</td>
<td>Massive, rare current-ripple cross-lamination</td>
<td>Root traces at the uppermost portion may penetrate the entire thickness.</td>
<td>Abandoned-channel fill ref.: Fisk 1944; Allen 1965; Bridge 2003, 2006; Kraus and Davis-Vollum 2004</td>
</tr>
<tr>
<td>7</td>
<td>Extensive tabular mudstone and siltstone</td>
<td>Mottled gray, light gray, or olive gray.</td>
<td>Planar lamination, low bioturbation (Fig. 8).</td>
<td>Rare preserved in situ tree trunks. Animal burrows.</td>
<td>Overbank flood deposit ref.: Coleman 1966; Miall 1996; Retallack 1997, 2001; McCarthy et al. 1999; Driese and Ober 2005</td>
</tr>
<tr>
<td>8</td>
<td>Confined laminated mudstone and siltstone</td>
<td>Dark gray, gray, light or olive gray.</td>
<td>Planar lamination, low bioturbation (Fig. 13E).</td>
<td>Root traces and drab-haloed root traces.</td>
<td>Floodplain lacustrine fill ref.: Coleman 1966; Miall 1996; Denko and Gastaldo 1996</td>
</tr>
<tr>
<td>10</td>
<td>Carbonaceous mudstone and siltstone</td>
<td>Dark gray to black.</td>
<td>Massive (Figs. 8, 15A).</td>
<td>None</td>
<td>Peat-swamp or backswamp deposit ref.: McCabe 1984; Diesel 1992; Miall 1996; Retallack 2001; Driese and Ober 2005</td>
</tr>
<tr>
<td>11</td>
<td>Coal</td>
<td>Black</td>
<td>Fractures. Sulfur Residue (Figs. 8, 12D, 15A).</td>
<td>Teredolites (Figs. 12C, 12D).</td>
<td>Peat swamp or and backswamp deposit ref.: Flores and Hanley 1984; Bromley 1984; Savrda 1991; Diesel 1992; Benton and Harper 1997; Bohacs and Suter 1997; Soil Survey Staff 1999; Birkeland 1999; Retallack 2001; Gingras et al. 2004</td>
</tr>
</tbody>
</table>
Fluvial Aggradation Cycle Sets

Fluvial aggradation cycle sets (FAC sets) are mappable meter-to-decimeter-thick aggregates of FACs. They are bounded by pedogenically modified surfaces of rooting and burrowing or are erosionally top-truncated. The subaerially exposed surfaces are commonly capped by coals and carbonaceous mudstones or siltstones. FAC sets are marked by an upward increase in the degree of rooting and disruption of original sedimentary structures. They are represented by either compound floodplain bedsets or multi-story channel sandstones that are both capped by relatively mature paleosols.

Figure 16 shows two examples of a FAC set. The first example (Fig. 16B) consists of six vertically stacked, paleosol-capped floodplain bedsets (or FACs). The uppermost paleosol is relatively well rooted with a thicker zone of disruption compared to other paleosols in the set. The upper paleosol is also capped by carbonaceous mudstone and coal (i.e., a coal zone). In the second example (Fig. 16C), two multi-story channel sandstones are capped by a well-rooted paleosol that is overlain by carbonaceous siltstone and coal.

A total of nine fluvial aggradation cycle sets (henceforth FAC sets) were identified in the studied interval. Their thickness varies from about 5 m to 10 m for the best representative sections with minimal truncation. They are more laterally extensive than FACs and, barring local erosion by overlying channel belts, can be correlated across each cross section, for a minimum of several kilometers (Figs. 17, 18). The coal zones that cap the bounding surface of FAC sets also aid in their correlation. Given that there are no pinch-outs of the FAC sets (excepting local top-truncation), it is very likely that FAC sets extend for distances that greatly exceed the study area, likely tens of kilometers and possibly greater. The 9 preserved FAC sets are named in ascending numerical order from bottom to top and are discussed in detail below.

FAC-Set 1.—A major erosional surface separates FAC-set 1 from the underlying marine shoreface (FA. 1) deposits (Figs. 8, 17, 20A). FAC-set 1 varies from 2 to 4 m in thickness and consists of amalgamated, multi-story channel sandstones (FA. 2), overlain by tide-influenced channel sandstones (FA. 3). This contact (Figs. 17, 20A) has been identified in several studies as a regional erosional surface (base of S1 in Fig. 3A) that extends for several kilometers beyond the study area (Li et al. 2010; Zhu et al. 2012; Li and Bhattacharya 2013). The juxtaposition of these channels and underlying marine shoreface deposits (Fig. 20A), across an intervening erosional surface represents a basinward shift in facies (Walker and Plint 1992; Posamentier et al. 1992; Posamentier and Allen 1999) accompanied by valley incision that led to truncation of underlying marine facies. This surface represents a sequence boundary (Posamentier and Vail 1988; Van Wagoner et al. 1990; Posamentier and Allen 1999) and marks the base of Sequence 1. FAC-set 1 is represented by the channel sandstones that were deposited after the valley incision phase.

The trace fossil *Gastrochaenolites* was identified at the sequence boundary, and *Ophiomorpha* and *Skolithos* were identified in the lower portions of the valley-fill sandstones in FAC-set 1 (Fig. 21A, B). *Gastrochaenolites* indicates firm-ground development and colonization by marine organisms after the erosion of unconsolidated or semiconsolidated marine substrate (MacEachern et al. 1992; Benton and Harper 1997). The trace-fossil suite also suggests the existence of short-lived marine or marginal-marine conditions before increased fluvial deposition (MacEachern et al. 1992; Bromley 1996). The existence of
marine or marginal-marine conditions indicates proximity of this part of the incised valley to the shoreline (MacEachern et al. 1992), thus presenting an example of a coastal-plain incised valley.

Paleocurrent data (Fig. 22A) indicate a north to northwest paleo-flow direction for all valley-fill sandstones in FAC-set 1. Based on the thickness of largescale inclined strata (2.8–3.2 m), which are interpreted to represent unit bars in the formative channel, paleo-flow depths of 3.6 to 4 m were estimated (Bridge 2003). Estimated width (using numerical methods), based on bankfull depth vary from 500 to 650 m, and flow velocity based on medium-grained sandstones and abundance of dune-scale cross bedding varied from 0.6 to 1.5 m/s. Based on these parameters, the estimated bankfull paleo-discharge \( (Q_w) \) for the rivers in the paleo-incised valley was estimated to have varied from 750 to 2500 m\(^3\)/s. (Table 5). This indicates trunk streams in a regional paleo-incised valley. In comparison to typical trunk stream depths in the Ferron Sandstone (~ 7 m), these are probably not trunk streams. Their relatively small scale suggests that they are tributary channels at the valley margin, which are linked to larger trunk streams farther north (Li et al. 2010; Li and Bhattacharya 2013) (Fig. 3A).

**FAC-Set 2.**—A major erosional surface separates FAC-set 1 from the overlying FAC-set 2, which is about 2 to 4 m thick (Figs. 17, 20B). This erosional surface truncates both FAC-set 1 and locally erodes into the underlying marine

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**Fig. 10.—**A) Uninterpreted and B) interpreted lateral-accretion sets in a channel sandstone draped by abandoned-channel fill. Yellow arrows indicate direction of accretion.
shoreface deposits. This surface has also been identified as a regional erosional surface (Li et al. 2010; Li and Bhattacharya 2013). It suggests another erosional event associated with valley incision, and provides support for the compound nature of the valley system, as identified by other researchers (Li et al. 2010; Li and Bhattacharya 2013).

Figure 20B shows this younger incised valley truncating underlying marine shoreface deposits and cutting out the deposits of FAC-set 1. Root traces in truncated sandstones of underlying FAC-set 1 (Fig. 21C) indicate plant colonization and vegetative growth due to subaerial exposure. FAC-set 1 therefore represents a preserved terrace. The resulting second incision was filled with
tide-influenced heterolithic strata capped by an organic-rich and burrowed siltstone (FA. 3) and comprises FAC-set 2. Paleocurrent data obtained from the heterolithic strata indicate a bimodal paleo-flow trend (Fig. 22B) suggesting landward tidal flow and shoreward river flow.

Paleo-flow depths of 1.6 to 1.8 m in the associated channel were estimated based on thicknesses of preserved channel stories. Estimated width (using numerical methods), based on paleo-flow depth, varies from 150 m to 170 m. Flow velocity, based on very fine- to fine-grained sandstones and abundance of dune-scale cross bedding, varied from 0.7 to 1.2 m/s. Based on these parameters, the estimated bankfull paleo-discharge \( Q_w \) for the river likely varied from 120 to 200 m\(^3\)/s (Table 5). The smaller-scale channels suggest a tributary system linked to larger trunk streams farther north.

**FAC-Set 3.**—FAC-set 2 is capped by FAC-set 3, which is a 3 to 5 meter-thick mudstone-prone unit (Figs. 17, 18). It consists of partially exposed isolated channel sandstones (FA. 2) encased in floodplain deposits (FA. 4, 7–11). FAC-set 3 is capped by coal zone 1 (CLZ 1), which is an extensive coal zone that has been identified and mapped in other parts of the Ferron Notom Delta (Zhu et al. 2012) (Fig. 3) and is thus regional in nature. Limited exposure of channel sandstones prevented the collection of paleocurrent data and consequent estimation of channel dimensions in this FAC set.

**FAC-Set 4.**—Overlying FAC-set 3 is FAC-set 4, which is about 2 to 3 m thick. It is mostly erosionally truncated by the overlying FAC-set 5 and is exposed only in the north study area (Fig. 17). It consists of a tide-influenced, heterolithic channel-fill (FA. 3) encased in floodplain mudstones (FA. 7, 9, 10). The heterolithic channel fill consists of south-dipping accretion sets. Paleocurrent data indicate east to northeast paleo-flow, and thus indicate laterally accreting channels. A paleo-flow depth of 2.1 m was estimated based on thicknesses of the preserved channel story. Estimated width (using numerical methods), based on paleo-flow depth, varies from 200 to 230 m. Flow velocity, based on very fine- to fine-grained sandstones and abundance of dune-scale cross bedding, varied from 0.8 to 1.2 m/s. Based on these parameters, the estimated bankfull paleo-discharge \( Q_w \) for the depositing river may have varied from 220 to 400 m\(^3\)/s (Table 5). These likely represent distributary channels on an upper delta plain.

**FAC-Set 5.**—FAC-set 5 is 4 to 10 m thick (Figs. 17, 18), and it consists of laterally extensive, amalgamated and locally multi-story channel sandstones (FA. 2) with attached crevasse-splay (FA. 4) and levee (FA. 5) deposits overlain by other floodplain deposits (FA. 7, 9–12). The paleosols that cap this unit represent the most developed paleosols observed in this study. These paleosols show well-developed horizons, significant ped structure development, high degree of rooting (R.I. = 5), and significant pedogenic clay accumulation. They are comparable to modern floodplain Alfisols (Type V; Table 3) that develop in 1 ka to 5 ka of subaerial exposure (Aslan and Autin 1998). These paleosols are capped by coal zone 2 (CLZ 5). The thickness of fully preserved channel stories varies from 3 to 6.5 m. Paleocurrent data (Fig. 22C) indicate a predominant north to northwest paleo-flow direction for the depositing river.

Estimated paleo-flow depth is based on the thicknesses of these preserved stories (3 to 6.5 m). Channel widths were estimated using numerical methods based on paleo-flow depth and preserved channel width from outcrop data.
Fig. 13.—Floodplain facies. A) Fining-upward crevasse-splay channel sandstone with an erosional base. B) Crevasse-splay facies showing two individual mudstone-capped sandstones representing flood-generated sedimentation units (or rhythmites). C) Levee facies flanking channel-fill sandstone. D) An expanded view of the area marked by red square in Part C showing rhythmites. E) Floodplain lake-fill facies with preserved tree stump and siderite concretions. F) Abandoned-channel fill preserved in channel thalweg (rock hammer for scale).
These width estimates vary from 200 to 1200 m. Flow velocity, based on medium- to coarse-grained sandstones and abundance of dune-scale cross bedding, varied from 0.6 to 1.5 m/s. Based on these parameters, the estimated bankfull paleo-discharge ($Q_{bw}$) for the depositing river may have varied from 350 to 4200 m$^3$/s (Table 5). The larger rivers are interpreted as trunk streams, although they do not appear to be incised valleys, and the increase in channel size may reflect an overall seaward migration of the fluvial system, with respect to the underlying FAC Set. The more mature paleosol that caps the FAC set may reflect a more prolonged period of subaerial exposure than in the underlying FACs and FAC sets.

**FAC-Sets 6 and 7.**—FAC-set 6 and FAC-set 7 are successive mudstone-prone units. The thicknesses of these FAC sets vary from 1 to 5 m (Figs. 17, 18). They consist of a series of well-exposed lens-shaped, isolated channel sandstones (FA. 2) that are encased in floodplain deposits (FA. 4–11). FAC-set 6 and 7 are capped by coal zones 3 and 4, respectively. Isolated channel sandstones are mostly single-story, and full story thicknesses vary from less than 1 to a maximum of 5 m. Paleocurrent data (Fig. 22D, E) show that paleo-flow direction varies for individual channel sandstones; however, they may indicate northeast, north, or northwest paleo-flow directions. Centimeter- to decimeter-thick, dipping accretion sets are observed in some isolated channel sandstones (Fig. 10), and their dip is often perpendicular to paleo-flow direction, thus indicating lateral accretion.

Paleo-flow depths of 1 to 5 m were estimated based on the thicknesses of preserved channel stories. Channel widths were estimated using numerical methods based on paleo-flow depth, preserved channel widths, and preserved lateral-accretion sets. These width estimates vary from 25 to 270 m. Flow velocity, based on very fine- to medium-grained sandstones and abundance of dune-scale cross bedding, varied from 0.5 to 1.4 m/s. Based on these parameters, the estimated bankfull paleo-discharge ($Q_{bw}$) for the depositing river may have varied from 25 to 800 m$^3$/s (Table 5). The larger channels may also represent trunk streams, although this system seems to record a backstep in facies, compared to the underlying FAC set.

**FAC-Set 8.**—FAC-set 8 is mostly erosionally top-truncated (Figs. 17, 18) by the overlying FAC-set 9, leaving less than 1 to 2 m of preserved strata, which
Table 3.—Characteristics of paleosol observed in the studied interval and their modern soil analogs.

<table>
<thead>
<tr>
<th>Paleosol Type</th>
<th>Characteristics</th>
<th>Typical Horizons</th>
<th>Modern Soil Analog</th>
</tr>
</thead>
<tbody>
<tr>
<td>Type I</td>
<td>Coal or carbonaceous shale.</td>
<td>O, O-C</td>
<td>Histosols</td>
</tr>
<tr>
<td>Type II</td>
<td>Poor horizon development. Poorly developed ped structures. Preserved original sedimentary structures. Low intensity of rooting. No pedogenic clay accumulation.</td>
<td>A, A-C</td>
<td>Entisols</td>
</tr>
<tr>
<td>Type III</td>
<td>Moderate horizon development. Incipient to moderate development of ped structures. Some pedogenic clay accumulation. Moderate intensity of rooting.</td>
<td>A-Btg-Bg-C</td>
<td>Inceptisols</td>
</tr>
<tr>
<td>Type IV</td>
<td>Poor to moderate horizon development. Incipient to moderate development of ped structures. Abundant slickensides. Some pedogenic clay accumulation often distorted by slickensides. Moderate intensity of rooting.</td>
<td>Btg-C, Btg Bsg-C</td>
<td>Vertisols</td>
</tr>
<tr>
<td>Type V</td>
<td>Well-developed horizons. Well-developed ped structures. Significant pedogenic clay accumulation. High intensity of rooting.</td>
<td>A, A, Btg, Bg-C</td>
<td>Alfisols</td>
</tr>
</tbody>
</table>

Paleosol Type Characteristics Typical Horizons Modern Soil Analog

Consist predominantly of finer-grained, muddy floodplain deposits (FA. 4, 7–11). FAC-set 8 is capped by coal zone 5 (CLZ 5).

FAC-Set 9.—FAC-set 9 consists of laterally extensive, amalgamated and multi-story channel sandstones (FA. 2), although channel-belt pinchouts can be observed locally (section 12-04, Fig. 17, and sections 11-02 and 11-06, Fig. 18). The thickness of preserved channel stories varies from 2 to 4 m (Figs. 17, 18). FAC-set 9 is capped by the fissile, dark-gray, parallel-laminated strata of the Blue Gate Shale with an intervening regional transgressive unconformity marked by a coarse-grained lag (Li et al. 2010). Fossil fish scales and strata of the Blue Gate Shale with an intervening regional transgressive unconformity marked by Thalassinoides fossils (and increased organic-matter preservation in the abandoned area. Marine trace fossils (Teredolites and Thalassinoides) and dinocysts above the coals show good evidence of marine transgressions as a result of either compactional subidence of the abandoned area (Ryer 1981) or eustatic sea-level rise (Bohacs and Suter 1997; Ryer and Anderson 2004).

These phases of soil development and peat accumulation suggest that the length of the floodplain and channel abandonment phase is longer than those that succeed FACs, and indicates a shift in locus of sedimentation (Bridge 1984; Kraus 1987; Atchley et al. 2004; Cleveland et al. 2007), as evidenced by the abandoned channel deposit, carbonaceous mudstones, and coal that cap the FAC set in Figure 16B and 16C. This length of abandonment is likely linked to significant avulsion.

Interpretation of FAC Sets

FAC sets are interpreted to record successive phases of FAC development (i.e., continuous floodplain and channel aggradation interrupted by short phases of abandonment and pedogenesis) that eventually culminate in periods of stability characterized by abandonment, exposure, and significant pedogenic development. Soils that form during this phase of abandonment show moderate horizon development, poorly to moderately developed ped structures, destruction of original sedimentary fabric, and some pedogenic clay accumulation. The entire FAC set shows upward increase in the intensity of rooting, which suggests extensive vegetative growth during abandonment. High root-abundance percentages observed by Nordt et al. (2011) in Upper Cretaceous to Paleocene floodplains support this observation. Paleocurrent data from channel sandstones indicate paleo-flow in a predominantly northeast direction (Fig. 22F). The lag may have truncated the top of FAC-set 9, resulting in incomplete preservation of the uppermost channel stories. Channel stories in the lower FACs typically include ripple cross-laminated sandstones, but these upper channel-fill facies are notably absent in FAC-set 9, consistent with top truncation. Also, the thickness of dune-scale cross sets is similar in channel fills to those seen in the lower thicker stories (e.g., FAC-set 5), which may indicate similar channel flow depths, given that the thickness of dune-scale cross sets scales to flow depth (e.g., LeClair and Bridge 2000; Bhattacharya and Tye 2004; Li et al. 2010).

Paleo-flow depth was estimated based on thicknesses of preserved channel stories (i.e., 2 to 4 m), but these could be somewhat larger given the lack of preserved upper-channel fill. Channel widths were estimated using numerical methods based on paleo-flow depth and preserved channel widths. These width estimates vary from 300 to 600 m. Flow velocity, based on medium- to coarse-grained sandstones and abundance of dune-scale cross bedding, varied from 0.65 to 1.65 m/s. Based on these parameters, the estimated bankfull paleo-discharge (Qw) for the depositing river may have varied from 350 to 1000 m³/s (Table 5). These are interpreted as trunk streams (considering the upper truncation). They likely reflect an overall seaward migration of the fluvial system with respect to the underlying FAC set.

Time Duration of FAC Sets

In order to estimate the approximate time of development for FAC sets, the most representative sets with the lowest (i.e., two) and highest (i.e., six)

Table 4.—Summary of observed stratal hierarchy in the Ferron Notom Sequence 1. Duration includes time represented by deposition and paleopedogenesis.

<table>
<thead>
<tr>
<th>Stratal Hierarchy</th>
<th>Observed Thickness (m)</th>
<th>Observed Lateral Extent (m)</th>
<th>Duration (yrs)</th>
</tr>
</thead>
<tbody>
<tr>
<td>FAC</td>
<td>0.5–6.5</td>
<td>10s–100s</td>
<td>25–2,000</td>
</tr>
<tr>
<td>FAC Set</td>
<td>5–10</td>
<td>100s–1,000</td>
<td>240–14,000</td>
</tr>
<tr>
<td>Fluvial</td>
<td>&gt; 10</td>
<td>1000s–10,000s</td>
<td>20,000–40,000</td>
</tr>
</tbody>
</table>

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numbers of FACs were identified (i.e., those not affected by significant truncation), and each correspond to a time duration of 50 to ≤ 12,000 years based on earlier estimated time duration of FACs. The characteristics of paleosols that developed after the abandonment that followed the deposition of FAC set are similar to those of modern Inceptisols (Type III; Table 3) and Vertisols (Type IV; Table 3), and typically represents a few hundreds to a thousand years (100 a to 1 ka) of pedogenic development (U.S. Soil Survey Staff 1999; Retallack 1988, 2001, Aslan and Autin 1999; Autin and Aslan 2001; Driese and Ober 2005).

To estimate the time of deposition of coal beds that cap bounding surfaces of FAC sets, a range of peat compaction ratios from 1.2:1 to 2.2:1 was used, based on estimation from field methods compiled by van Asselen et al. (2009) and consistent with observations made by Nadon (1998). The underlying assumption in using these ratios is that compaction of peat occurred early and at very shallow depths (McCabe 1987; Nadon 1998). Similar decompaction ratios have also been estimated by Rajchl and Uličný (2005) in a similar sedimentary succession from the Oligo-Miocene Ohre Rift Basin in the Czech Republic.
FIG. 17.—Two-dimensional stratigraphic cross section of the Ferron Notom in North Sweetwater showing the stacking of fluvial aggradation cycles and fluvial aggradation cycle sets into three fluvial sequences (sequence 1C, 1B, and 1A). See cross section outline in Figure 21, and see legend in Figure 7.
Using an average peat accumulation rate of 2.5 mm/yr, which has been reported for modern tropical deltas (McCabe 1984; Emery and Myers 1996; Page et al. 2010), estimates in Table 6 show that the peat accumulation time for coal zones in this study vary from 100 to 600 years. Based on these estimates, FAC sets in this interval likely represent no more than 14 ka (i.e., 250 a to $\approx$ 14 ka).

This estimate shows that these FAC sets are comparable in duration (i.e., 1 ka to $> 10$ ka) to the macroscale stratal hierarchy of Kraus and Aslan (1999). They are also similar in duration (i.e., 10 a to $> 100$ ka) to the FAC sets described by Atchley et al. (2004), Prochnow et al. (2006), and Cleveland et al. (2007).

**Fluvial Sequences**

In the studied interval, fluvial sequences are mappable decameter-scale units that consist of multiple vertically stacked FAC sets bounded by major unconformities (erosional or nondepositional). Similar to marine sequences, fluvial sequences can consist of systems tracts, which include: falling-stage, lowstand, transgressive, and highstand (Van Wagoner et al. 1988; Shanley and McCabe 1993, 1994; Miall 1991; Wright and Marriott 1993; Posamentier and Allen 1999). However, FAC sets are not necessarily equivalent to systems tracts; a single FAC set may represent parts of two systems tracts. Likewise, a single FAC set or multiple FAC sets may represent a single systems tract (cf. Atchley et al. 2004). The characteristics of each FAC set and its relationship to underlying and overlying strata determine its systems-tract designation.

The depositional history of the units in this study is illustrated in cross sections to show the different hierarchies of strata (Figs. 17, 18) and on Wheeler diagrams (Fig. 23A, B) to show stratigraphic relationships in time and space. The marine strata equivalent to this fluvial succession are not preserved due to modern erosion, hence direct comparisons could not be made. However, the lower Ferron Notom sequences contains a well-studied marine component that allows us to compare the nature of marine versus nonmarine cyclicity within a single genetically related clastic wedge deposited in a short (600 ka) time period.

Also, palynological analysis (Akyuz et al. 2015) indicates swampy everwet conditions in the studied interval, and shows marine dinocysts at four stratigraphic levels that coincide with previously identified flooding surfaces that mark FAC set boundaries. Marine dinocysts were also relatively abundant in incised-valley fills. Marine trace fossils associated with coals at these FAC set boundaries and in incised-valley fills also confirm marine influence (Bromley et al. 1984; Savrda 1991; MacEachern and Pemberton 1992; Benton and Harper 1997; Bohacs and Suter 1997; Gingras et al. 2004).

The assignment of FAC sets to systems tracts and sequences are based on: (i) interpreted changes in process and environments in between successive FACs and FAC sets, (ii) palynological analysis, (iii) comparison to fluvial sequence-stratigraphic models (Miall 1991, 1996; Wright and Marriott 1993), and (iv) similar field observations from other studies (Olsen et al. 1995; Yoshida et al. 1996; Plint et al. 2001; Fanti and Catuneanu 2010).

We interpret the valley incision preceding FAC-set 1, and the erosion of underlying marine shoreface deposits to be a result of base-level fall, forming the equivalent to a falling-stage systems tract. Valley filling was marked by moderate stream discharge and deposition of amalgamated multi-story channel sandstones and tide-influenced channel sandstones of FAC-set 1. The relatively small scale of channels ($\approx 4$ m) suggests that these may be tributary channels at the valley margin. Thicker stories have been documented in exposures of this compound valley farther north (Li et al. 2010; Li and Bhattacharya 2013). FAC-set 1 is interpreted as a lowstand to early transgressive systems tract.

Another phase of erosion and valley incision and terrace development, as a result of base-level fall, resulted in truncation of the underlying incised-valley fill (FAC-set 1) and marine shoreface deposits. This represents
another falling-stage systems tract. Subsequent base-level rise led to deposition of tide-influenced heterolithic strata (FAC-set 2), which is interpreted as representing the lowstand to early transgressive systems tract. Zhu et al. (2012) interpreted the organic-rich and burrowed capping surface of FAC-set 2 as a marine flooding surface, which is also confirmed by relative abundance of marine dinocysts (Akyuz et al. 2015). Alternatively, this can be interpreted entirely as an early transgressive systems tract (Dalrymple et al. 1994). In that case, the amalgamated channel deposits that would typically represent the lowstand systems tract were probably not deposited at this location. Deposition of FAC-set 2 concludes the incised-valley deposition phase (Fig. 17). Channels are relatively small (<2 m), probably indicative of tributary systems.

FAC-set 1 and FAC-set 2 represent separate incised-valley deposits in a compound incised-valley system, where FAC-set 1 represents the first cut-and-fill phase and FAC-set 2 represents the second phase of truncation and filling, and subsequent marine transgression as evidenced by marine trace fossils and marine dinocysts. Multiple cut-and-fill episodes produced a diachronous erosional surface or a regional composite scour surface (RCS) (Holbrook and Bhattacharya 2012).

Continued base-level rise was accompanied by deposition above and outside the compound incised valley. This was marked by increased floodplain deposition and preservation, and deposition of smaller-scale distributary-channel sandstones with limited lateral continuity (FAC-set 3). FAC-set 3 is interpreted as a late transgressive systems tract. Abundant marine trace fossils and marine dinocysts found in coal zone 1, which caps FAC-set 3, indicate marine transgression due to continued rise in base level, and may represent a maximum flooding surface (Akyuz et al. 2015). Zhu et al. (2012), based on more regional correlations to the north, also interpreted this surface as a marine flooding surface.

The signature of this marine transgression is still evident in the overlying FAC-set 4, where evidence of marine influence is also observed in the laterally accreting heterolithic channel fill (Fig. 17). FAC-set 4 is interpreted as an early highstand systems tract and marks the beginning of the “turn-around” or decrease in the rate of base-level rise. Decrease in base-level rise continued through the deposition of FAC-set 5, which eroded most of the early highstand deposits of FAC-set 4 and was characterized by deposition of laterally extensive, amalgamated channel sandstones. This reduced floodplain preservation due to continued reworking and removal of floodplain sediments by laterally migrating channels. The scale of channels (up to 6 m thick) suggests that they are trunk channels, although there is no evidence that they were significantly incised as another valley system. FAC-set 5 is thus interpreted as the late highstand systems tract.

The subsequent falling-stage, lowstand, and possibly early transgressive systems tracts are not represented in the succession, as might be indicated by incised valleys. Erosion and deposition that accompanied these missing systems tracts probably occurred in distal parts of the delta, down depositional dip away from the study area. However, FAC-set 5 is capped by the most developed and mature paleosol, characterized by well-developed horizons, significant development of ped structure, high degree of rooting (R.I. = 5), and significant pedogenic clay accumulation, which suggests a sequence boundary (Van Wagoner et al. 1990; McCarthy and Plint 1998). Coal zone 2, which caps the paleosol, suggests initial marine transgression. FAC-set 5 is directly overlain by FAC-sets 6 and 7, which are characterized by increased preservation of floodplain deposits, and deposition of smaller isolated channel sandstones indicating a backstep and return to more distributary channels from the trunk streams of FAC-set 5. This interval is interpreted as another transgressive systems tract. Abundant marine trace fossils, and marine dinocysts found in the extensive coal zone 4 that cap FAC-set 7 indicate marine transgression, and proximity to a shoreline, and it is interpreted as the fluvial equivalent of a maximum flooding surface.

FAC-set 8 is interpreted as an early highstand systems tract and marks the beginning of decrease in base-level rise. Most of FAC-set 8 was eroded during the subsequent late highstand systems tract, which was characterized by continued decrease in base-level rise which led to deposition of laterally extensive amalgamated channel sandstones and reduced floodplain preservation (FAC-set 9). Channels do not exceed about 4 m in thickness, but there is significant top truncation, so these could represent top-eroded trunk channels. Certainly
the high degree of amalgamation suggests lowered accommodation, although pinchouts of the top channel belts are observed locally. The contact between deposits of this late highstand systems tract (FAC-set 9) and the overlying Blue Gate Shale is both a regional unconformity, which is likely tectonic in origin, and a regional transgressive surface of erosion, as indicated by the coarse-grained sandstone lag and the preservation of marine fish scales at the contact (Fig. 17; Li et al. 2010; Zhu et al. 2012).

The above interpretation reveals three separate depositional sequences (Figs. 17, 18, 23A, B) in what was formerly interpreted as a single depositional sequence (i.e., Sequence 1). These are: Sequence 1C (oldest), Sequence 1B, and Sequence 1A (youngest). The first and oldest sequence (Sequence 1C) comprises falling stage, lowstand, and transgressive systems tracts. The second sequence (Sequence 1B) comprises falling stage, lowstand, transgressive, and highstand systems tracts. The third and youngest sequence 1A comprises a transgressive and highstand systems tract. Both Sequences 1C and 1B can be classified as simple sequences that comprise a sequence set, whereas Sequence 1A represents a “stand-alone” simple sequence. More broadly, the originally defined Sequence 1 of Zhu et al. (2012) is not a simple sequence, but comprises a possible sequence set and a simple sequence.

Figure 23A and 23B show Wheeler diagrams for the entire sequence, where FACs and FAC sets are separated by small time gaps. The small time gaps represent periods of paleosol development or erosional truncation. The major time gaps represent periods of erosion (vacuities) or nondeposition (hiati), which are often regional and are interpreted as sequence boundaries. This diagram also shows that a significant portion of time in the overall development of this sequence is represented by erosion, nondeposition, and paleosol development.

**CONTROLS ON CYCLICITY**

The along-strike variation in the distribution of FACs in the studied interval and the observation that many FACs do not extend across the study area suggest that the factors that control their distribution are largely internal to the
Incised-valley trace fossils and valley terrace relic. A) *Gastrochaenolites* (Ga) at the erosional contact between marine shoreface sandstone and fluvial incised-valley fill (i.e., FAC set-1) suggests firmground development and colonization by marine organisms after the erosion of unconsolidated or semiconsolidated marine substrate. B) *Ophiomorpha* (Op) in the lower portion of the incised-valley fill (FAC-set 1) suggests short-lived marine or marginal-marine conditions before increased fluvial deposition. C) Root traces (identified with arrows) in truncated sandstones of FAC-set 1 suggest terrace development and plant colonization.

Rose diagrams showing paleocurrent directions for channel fills in the respective FAC sets. Data for FAC-set 4 and FAC-set 8 are unavailable due to limited exposure.
Based on the observations of Nadon (1998).

Based on the methods of Bridge and Mackey (1993) and Fielding and Crane (1987).

1.3 300

5 200

5 200

2.0

5

3.6

(e.g., Kraus and Aslan 1999; Atchley et al. 2004; Prochnow et al. 2006; Bhattacharya 2013) attributed repeated incision and erosion in the basal composite sequences to the shortest higher-frequency Milankovitch-scale cycles (≈ 100 ka), and simple sequences in a forced-regressive sequence set to the shortest higher-frequency Milankovitch-scale cycles (< 30 ka). Based on duration estimates (240 a to ≤ 14 ka) and number of FAC sets (9 FAC sets), the estimated time duration for sequence 1 (including 1A, 1B, and 1C) is 2 to ≤ 126 ka. The time estimate of ~ 100 ka by Zhu et al. (2012) falls within the upper end of this range, and this indicates some level of agreement. Furthermore, Li and Bhattacharya (2013) attributed repeated incision and erosion in the basal composite incised valley of Sequence 1 to stepped falls in relative sea level (i.e., stepped, forced regressions) followed by sea-level rises. Each “fall and rise” phase was also correlated with higher-frequency Milankovitch-scale cycles of less than 30 ka. Therefore, the interpreted sequences in this studied interval (Sequence 1C, 1B, and 1A) likely correlate to the shortest, higher-frequency Milankovitch-scale cycles (< 20 ka, ~ 40 ka) similar to that of Li and Bhattacharya (2013). Observations from this study suggest that sequence-boundary development spanned 1 to 5 ka. These sequences are similar in scale in terms of thickness (i.e., > 10 m) and duration (> 10⁷ years) to the upper end of previously identified macrostratal hierarchy of Kraus and Aslan (1999), which has been attributed to a combination of allogenic and autogenic processes. However, based on earlier observations in previous studies and Cleveland et al. (2007) suggests that a combination of autogenic and allogenic processes build FAC sets. These may include base-level changes and regional avulsions (i.e., covering several kilometers).

Furthermore, the scale of cyclicity in FACs is comparable to those estimated for marine parasequences in the Ferron Notom (about 10,000 to 15,000 years), which suggests a possible link between FAC-set development in the coastal plain and down-dip marine parasequence development. Marine trace fossils and marine dinocysts associated with FAC-set boundaries suggest that they are coastal-plain extensions of parasequence boundaries. However, there is no downdip marine section to link with this sequence, and the best inference we can make is from older sequences in the Ferron Notom succession that have linked fluvial and marine strata. Nonetheless, it is very likely that the development of FAC sets is influenced by base-level changes (allogenic process), which may have influenced autogenic processes (e.g., avulsion) in the alluvial plain.

TABLE 5.—Estimated channel-belt dimensions and corresponding paleo-discharge estimates.

<table>
<thead>
<tr>
<th>FAC-set</th>
<th>Paleo-Channel Depth (m)</th>
<th>Estimated Width (m) from Empirical Methods*</th>
<th>Estimated Width (m) from Outcrop Data</th>
<th>Flow Velocity** (m/sec)</th>
<th>Discharge (Q) (m³/sec)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>3.6-4.0</td>
<td>500-650</td>
<td>-</td>
<td>0.60-1.50</td>
<td>750-2500</td>
</tr>
<tr>
<td>2</td>
<td>1.6-1.8</td>
<td>150-170</td>
<td>-</td>
<td>0.75-1.20</td>
<td>120-200</td>
</tr>
<tr>
<td>4</td>
<td>2.1***</td>
<td>200-230</td>
<td>-</td>
<td>0.80-1.20</td>
<td>220-400</td>
</tr>
<tr>
<td>5</td>
<td>3.0-6.5</td>
<td>200-1200</td>
<td>200-120</td>
<td>0.60-1.50</td>
<td>350-4200</td>
</tr>
<tr>
<td>6 &amp; 7</td>
<td>1.4-5.0</td>
<td>25-270</td>
<td>25-230</td>
<td>0.50-1.40</td>
<td>25-800</td>
</tr>
<tr>
<td>9</td>
<td>2.0-4.0</td>
<td>300-360</td>
<td>360-360</td>
<td>0.65-1.65</td>
<td>350-1000</td>
</tr>
</tbody>
</table>

* Based on the methods of Bridge and Mackey (1993) and Fielding and Crane (1987).

** Based on the three-dimensional bedform phase diagram of Rubin and McCulloch (1980).

*** Only one channel belt was estimated in FAC-set 4 due to outcrop limitations.

Furthermore, examination of the regional cross section (Fig. 3) shows both confined fluvial systems (in valleys associated with Sequences 1 and 2) and unconfined fluvial systems, especially in the upper part of sequence 1, similar to those described in this study. Hajek et al. (2010) suggest that clusters of unconfined channel belts may represent avulsion clusters, reflecting a tendency to development of local avulsions. Examination of Figure 17 shows that the unconfined channel belts in FAC-sets 6 and 7 (e.g., sections 11-18 to 11-15) are intimately associated with levee and crevasse-splay deposits (FA. 4 and 5). The association of channel-belt, levee, and splay deposits have also been used as a criterion to identify floodplains in which local avulsion is a dominant process (Jones and Hajek 2007). There is good evidence that the unconfined channel belts were dominated by frequent and local avulsions. This is also compatible with the interpretation of the smaller unconfined channel belts forming on a delta plain.

In contrast, FAC sets are laterally extensive and can be easily correlated for hundreds of meters to several kilometers beyond the extent of the study area (Fig. 23A, B). Similar studies have shown that FAC sets can be correlated over areas of several tens of kilometers, but few studies have attempted correlation across larger distances (Cleveland et al. 2007). This suggests that the factors which control the development of FAC sets are regional with significant influence across the alluvial plain. The similarity in time of development (100s to 1000s of years) to the lower end of comparable stratigraphic hierarchies (e.g., Kraus and Aslan 1999; Atchley et al. 2004; Prochnow et al. 2006; Cleveland et al. 2007) suggests that a combination of autogenic and allogenic processes build FAC sets. These may include base-level changes and regional avulsions (i.e., covering several kilometers).

Furthermore, the scale of cyclicity in FACs is comparable to those estimated for marine parasequences in the Ferron Notom (about 10,000 to 15,000 years), which suggests a possible link between FAC-set development in the coastal plain and down-dip marine parasequence development. Marine trace fossils and marine dinocysts associated with FAC-set boundaries suggest that they are coastal-plain extensions of parasequence boundaries. However, there is no downdip marine section to link with this sequence, and the best inference we can make is from older sequences in the Ferron Notom succession that have linked fluvial and marine strata. Nonetheless, it is very likely that the development of FAC sets is influenced by base-level changes (allogenic process), which may have influenced autogenic processes (e.g., avulsion) in the alluvial plain.

Zhu et al. (2012) attributed individual sequences in the Ferron Notom to relatively high-frequency Milankovitch-scale sequences (~100 ka), and simple sequences in a forced-regressive sequence set to the shortest higher-frequency Milankovitch-scale cycles (<30 ka). Based on duration estimates (240 a to ≤ 14 ka) and number of FAC sets (9 FAC sets), the estimated time duration for sequence 1 (including 1A, 1B, and 1C) is 2 to ≤ 126 ka. The time estimate of ~ 100 ka by Zhu et al. (2012) falls within the upper end of this range, and this indicates some level of agreement. Furthermore, Li and Bhattacharya (2013) attributed repeated incision and erosion in the basal composite incised valley of Sequence 1 to stepped falls in relative sea level (i.e., stepped, forced regressions) followed by sea-level rises. Each “fall and rise” phase was also correlated with higher-frequency Milankovitch-scale cycles of less than 30 ka. Therefore, the interpreted sequences in this studied interval (Sequence 1C, 1B, and 1A) likely correlate to the shortest, higher-frequency Milankovitch-scale cycles (~20 ka, ~40 ka) similar to that of Li and Bhattacharya (2013). Observations from this study suggest that sequence-boundary development spanned 1 to 5 ka. These sequences are similar in scale in terms of thickness (i.e., >10 m) and duration (>10^7 years) to the upper end of previously identified macrostratal hierarchy of Kraus and Aslan (1999), which has been attributed to a combination of allogenic and autogenic processes. However, based on earlier observations in previous studies and

TABLE 6.—Estimated duration of peat accumulation for coal bed and coal zones that cap FAC sets.

<table>
<thead>
<tr>
<th>Coal Zone</th>
<th>Associated FAC-SET</th>
<th>Coal Bed–Coal Zone Thickness (mm)</th>
<th>Decompacted Thickness* (x 1.2-2.2 mm)</th>
<th>Accumulation Period (years) (rate = 2.5 mm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>3</td>
<td>300–600</td>
<td>360–1320</td>
<td>144–528</td>
</tr>
<tr>
<td>2</td>
<td>5</td>
<td>200–600</td>
<td>240–1320</td>
<td>96–528</td>
</tr>
<tr>
<td>3</td>
<td>6</td>
<td>300–700</td>
<td>360–1540</td>
<td>144–616</td>
</tr>
<tr>
<td>4</td>
<td>7</td>
<td>250–700</td>
<td>300–1540</td>
<td>120–616</td>
</tr>
<tr>
<td>5</td>
<td>8</td>
<td>200–450</td>
<td>240–990</td>
<td>96–396</td>
</tr>
</tbody>
</table>

* Based on the observations of Nadon (1998).
earlier interpretations in this study, the dominant process in this case appears to be allogenic, and specifically base-level rise and fall.

Fielding (2011) recognized tectonic effects in unconformities in the Ferron Notom succession farther south of the study area. However, the uppermost part of the Ferron Notom in this study, which is equivalent to the "upper member" in the Fielding (2011) study, appears to have been unaffected by tectonic movements. In addition, the scale of the studied interval (~30 m) and its relatively short-duration sequence of strata may make detection of relatively large-scale and long-duration tectonic influences difficult.

Numerous previous estimates suggest that the paleo-rivers in the Ferron Notom wedge flowed over relatively steep slopes on the order of 10° (Li et al. 2010; Bhattacharyya et al. 2015). Given a maximum trunk channel depth of about 7 m, backwater effects would be expected to propagate landward from a few kilometers up to about 20 km. Bayline limits (representing the tidal range divided by the slope) would be even less, probably only a few kilometers. The regional dip cross section extends over about 30 km, and as a consequence, the nonmarine Ferron systems should alternate between being more or less affected by backwater effects. Also, evidence from paleosol types and stratal geometries suggests that there were small differences between trunk and tributary channel slopes, but it seems likely that the absolute gradient differences would have been small. If not, then the character of the "upland" paleosols would have been fundamentally different from those described here (in general more oxidized). The observations of dinocysts and local horizons of marine bioturbation support the idea that downstream sea-level controls may be more important than climate controls and tectonic controls, and that shorelines were within a few tens of kilometers of the study area. The palynological analysis (Akyuz et al. 2015) of spores and pollen also support an ever-wet environment, with little evidence for major changes in climate, such as wetting and drying cycles. If present, climate cycles might cause changes in the hydrologic cycles and correspondingly cause stratigraphic changes in river discharge through time (larger rivers reflecting wetter conditions, smaller rivers in drier times), but in this case, the changes in fluvial story thickness and stacking are more likely...
FIG. 23—Continued.
due to high-frequency base-level changes driven by eustatic sea level, rather than by climate or tectonics, as has been theorized in previous papers (e.g., Zhu et al. 2012; Li and Bhattacharya 2014).

CONCLUSIONS
A detailed outcrop characterization of what was originally interpreted as a single 100 ka, 4th- or 5th-order depositional sequence, consisting of three broadly defined parasequences reveals 11 depositional facies. These facies, in turn, combine to build at least 33 distinct fluvial aggradation cycles (FACs), 9 fluvial aggradation cycle sets (FAC sets), and 3 fluvial sequences.

Fluvial aggradation cycles (FACs) in the studied interval are represented by floodplain bedsets or individual channel stories that are capped by relatively immature paleosols or erosional surfaces. They may extend for tens to hundreds of meters. FACs represent periodic flood events or channel depositional events characterized by abandonment, subaerial exposure, and incipient pedogenesis, and they span not more than 2 ka. Fluvial aggradation cycle sets (FAC sets) are aggregates of FACs that are capped by relatively mature paleosols or erosional surfaces. They are represented by compound floodplain bedsets or multistory channel sandstones that are bounded by relatively mature paleosols, which are capped by coals and carbonate strata. They extend for hundreds of meters to several kilometers. FAC sets record successive phases of FAC development that eventually culminate in relatively long periods of stability characterized by abandonment, exposure, and moderate pedogenetic development, and they span not more than 14 ka.

Detailed facies description, correlation of FACs and FAC sets, and time-stratigraphic analysis reveal three high-frequency Milankovitch-scale (~20 ka, ~40 ka) fluvial depositional sequences that consist of complete or incomplete suites of depositional systems tracts. The oldest, Sequence 1C, truncates underlying marine shelf-form sandstones and consists of falling-stage, lowstand, and transgressive systems tracts. Sequence 1B truncates underlying Sequence 1C and underlying marine shelf-form sandstones, forming a compound erosional surface. Sequence 1B consists of falling-stage, lowstand, transgressive, and highstand systems tracts. The youngest Sequence 1A consists of transgressive and highstand systems tracts.

The development of FACs is attributed to mainly autogenic processes in the alluvial plain, while the development of FAC sets is attributed to a combination of autogenic and allocogenic processes in the alluvial plain. The scale of cyclicity of FAC sets is also similar to those estimated for marine parasequences in the alluvial plain, while the development of FAC sets is attributed to a combination of transgressive and highstand systems tracts. The youngest Sequence 1A compound erosional surface. Sequence 1B consists of falling-stage, lowstand, transgressive, and highstand systems tracts. The youngest Sequence 1A consists of transgressive and highstand systems tracts.

The development of FAC sets and sequences likely represents a dominance of downstream controls related to high-frequency sea-level changes, possibly linked to Milankovitch cycles, as previously theorized.

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