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E.A. Hajek1,†, P.L. Heller2, and E.L. Schur3
1Department of Geosciences, Pennsylvania State University, 511 Deike Building, University Park, Pennsylvania 16802, USA
2Department of Geology and Geophysics, University of Wyoming, 1000 E. University Avenue, Laramie, Wyoming 82071, USA
3Department of Geology, Macalester College, 1600 Grand Avenue, St. Paul, Minnesota 55104, USA

ABSTRACT

Internally generated (autogenic) sedimentary processes can obscure signals of tectonic movements, climate conditions, or sea-level fluctuations in alluvial basins. The stratigraphic effects of autogenic dynamics have been considered negligible on time scales of 10^4–10^6 yr. However, recent physical and numerical models have shown that over basin-filling time scales, self-organization in sedimentary systems can produce stratigraphic patterns similar to those resulting from changing basin boundary conditions. Here, we present the first field-based test for autogenic control on stratigraphy in an ancient alluvial basin. The Ferris Formation (Upper Cretaceous–Paleogene, Hanna Basin, Wyoming) is composed of clusters of closely spaced channel deposits separated by intervals dominated by overbank material—a pattern that has been proposed as an example of a potentially autogenic stratigraphy. In order to evaluate controls on Ferris Formation stratigraphy, spatial patterns of key channel properties were analyzed. These variables (including sand-body dimensions, paleoflow depth, maximum clast size, paleocurrent direction, and sediment provenance) were chosen because they can change in response to climate, tectonic, or sea-level forcing and are commonly measurable in ancient alluvial successions. No clear statistical trends were detected in the measured variables, which suggests that external forces did not likely control stratigraphic organization in the unit. This study demonstrates that autogenic dynamics in natural, field-scale systems can produce organized stratigraphic patterns over much larger spatial and temporal scales than is typically presumed. This finding emphasizes the need for further understanding of autogenic stratigraphy and greater care interpreting climate, tectonic, and eustatic changes from alluvial basins.

INTRODUCTION

Alluvial strata contain a primary record of past surface conditions on Earth, including information about mountain building, basin formation, climate conditions, and sea-level changes. The stratigraphic arrangement of channel-belt deposits in continental strata (alluvial architecture) can be used to interpret changes in patterns and rates of subsidence, sediment and water supply, and base level in ancient successions. Comparing the Ferris Formation stratigraphy to the overall stratigraphic pattern observed in the succession. Comparison between the Ferris Formation and stratigraphy produced in a physical experiment at the University of Minnesota (Sheets et al., 2007) is provocative, suggesting that clustering could result entirely from long time-scale autogenic organization in Ferris rivers (Hajek et al., 2010). However, the statistical similarity of channel patterns between the field and experiment is insufficient by itself to establish causative relationships between stratigraphic patterns and self-organization in an ancient fluvial system.

Here, we test for autogenic controls on channel clustering in the Ferris Formation by evaluating the ways in which channel variables that are likely sensitive to extrinsic controls relate to the overall stratigraphic pattern observed in the succession. Our hypothesis is that if changes in basin boundary conditions dominated the development of Ferris Formation alluvial architecture, there should be spatial relationships between measured channel variables and the clustered pattern of sand bodies within the basin. Conversely, no demonstrable correlation between key channel variables and channel locations within the study area indicates that Ferris Formation stratigraphy cannot be attributed to measurable external controls and the role of autogenic processes cannot be excluded.

To test this hypothesis, spatial data on six channel properties were collected across the study area. These include sand-body dimensions...
(width and thickness), maximum grain size, paleoflow depth, paleocurrent direction, sediment provenance, and overbank deposit character (e.g., erosion surfaces and paleosol development). As discussed later herein, these variables were chosen for their potential to indicate changes in river deposits due to climate, tectonic, and relative sea-level changes. The spatial distribution of these variables is evaluated across the study area as a function of both their association with specific channel-belt clusters and their overall stratigraphic position within the section.

Physical stratigraphic patterns, such as described here, are often the sole measureable record of paleolandscape conditions, and are therefore critical for interpreting past climate and tectonic changes. Ancient stratigraphy also provides a basis for understanding landscape evolution over time scales unobservable in modern and Holocene systems, and aids understanding and prediction of infrequent but potentially catastrophic hazards, including river avulsion. For these reasons, improved ability to identify autogenic stratigraphy is important for interpreting the sedimentary record.

**AUTOGENIC DYNAMICS OF RIVER AVULSION**

In systems where sediment deposition is focused along channels that aggrade or prograde more rapidly than floodplains, channels are driven to unstable configurations where they are perched (or "superelveled"; cf. Mohrig et al., 2000) above the surrounding basin (e.g., Martin et al., 2009; Mohrig et al., 2000) and/or develop steeper slopes away from the channel (across the floodplain) than down the channel (e.g., Tornqvist and Bridge, 2002). With continued channel aggradation, the likelihood increases that flow will avulse to a new position across the floodplain (e.g., Jones and Schumm, 1999; Slingerland and Smith, 2004). Channels that are at or near a critical superelvelation or slope threshold can be "triggered" to avulse by events such as a large flood, channel blockage, or earthquake (e.g., Gibling et al., 2010; Slingerland and Smith, 2004). Over geological time scales, avulsion triggers are ubiquitous, so avulsion frequency is dominantly controlled by the rate of channel aggradation/progradation relative to floodplain deposition (e.g., Ashworth et al., 2004; Bryant et al., 1995; Jerolmack and Mohrig, 2007; Reitz et al., 2010). Consequently, alluvial architecture studies in ancient settings often focus on understanding how basin boundary conditions, such as aggradation rates (set by sediment-supply rate and subsidence/base-level changes), influence basin-scale sandstone distributions (e.g., Allen, 1978; Bridge and Leeder, 1979; Heller and Paola, 1996; Kraus, 2002; Leeder, 1978; Sheets et al., 2002).

Channel avulsion is a quintessentially autogenic process that will occur both under steady and varying input conditions. Although spatial-temporal avulsion patterns have long been presumed to be quasi-random (e.g., Allen, 1978; Heller and Paola, 1996; Leeder, 1978), several numerical and physical modeling studies have shown that nonrandom avulsion patterns can arise spontaneously under constant boundary conditions (e.g., Jerolmack and Paola, 2007; Mackey and Bridge, 1995; Reitz et al., 2010; Sheets et al., 2007; Wang et al., 2011). These patterns can appear similar to structures produced by changes in basin subsidence, sediment supply, and incision/deposition cycles (Hajek et al., 2010). If avulsion processes can produce nonrandom stratigraphic patterns that mimic accepted indicators of changing boundary conditions, then autogenic avulsion patterns must first be ruled out in order to reliably interpret other signals.

**FERRIS FORMATION STUDY AREA**

The Upper Cretaceous–Paleogene Ferris Formation provides an exceptional opportunity to study the ways in which preserved river characteristics correlate to alluvial architecture. The field site in the northern Hanna Basin, Wyoming (Fig. 1), includes fluvial basin fill that has been rotated to a dip of ~80°, exposing a cross section of a Laramide basin across the present-day land surface (Fig. 2). This exposure affords direct access to hundreds of fluvial sand bodies where both their spatial distribution and sedimentologic properties can be mapped and measured.

The Ferris Formation is underlain by the marginal marine Medicine Bow Formation, which was deposited during the Maastrichtian as the Western Interior Seaway retreated generally to the east (Lillegraven and Ostresh, 1990). The base of the Ferris Formation is defined lithostratigraphically by the first appearance of gravels and is composed of fluvial sedimentary rocks deposited during uplift of the Ferris and Seminole Mountains directly north of the study area (Bowen, 1918; Ryan, 1977; Fig. 1). Despite being mere kilometers from some of the world’s richest Paleocene fossil localities (Eberle, 1995; Eberle and Lillegraven, 1998a, 1998b; Higgins, 2003; Lillegraven and Eberle, 1999), the Ferris Formation in the study area is devoid of preserved vertebrate remains and consequently lacks a refined chronostratigraphic framework. Based simply on the stratigraphic thickness of...
In Maastrichtian time, the broad Western Interior Basin was partitioned into smaller, regional sedimentary basins bounded by Laramide basement-cored uplifts (e.g., DeCelles, 2004; Dickinson, 2004; Dickinson et al., 1988). During this time, the Hanna Basin and neighboring Carbon Basin formed along the southern overthrust margin of a series of WNW-ESE-trending Laramide uplifts (Ferris, Seminoe, and Shirley Mountains and Freezout Hills; Fig. 1). Throughout the Late Cretaceous and early Paleocene, these ranges were presumably similar in size and vertical relief to other Laramide structures within the region, e.g., Wind River Range, Bighorn Mountains, which stood more than 2 km above surrounding basins (Dettman and Lohmann, 2000; Fan and Dettman, 2009). We identified the first appearance of basement (granite) clasts in the Ferris Formation more than 600 m below the base of the study area, suggesting that Ferris, Seminoe, Shirley, and Freezout ranges were well developed and fully unroofed long before the sediments in the study area were deposited.

Late Cretaceous through early Paleocene climate was generally warm and subtropical in south-central Wyoming and was characterized by strongly seasonal (i.e., monsoonal) precipitation patterns (e.g., Fricke et al., 2010; Sewall and Sloan, 2006). Regional paleoclimate models and proxy reconstructions suggest that the greater Green River, Wind River, and Powder River Basins (all within ~200 km of the Hanna Basin) had mean annual temperatures in the range ~10–25 °C and mean annual precipitation of ~40–150 cm (summarized in Sewall and Sloan, 2006), with heaviest precipitation occurring during the summer months when moisture was drawn from the Mississippi Embayment northwest into the Cordilleran highlands (Fricke et al., 2010; Sewall and Sloan, 2006). Additionally, stable isotope data suggest that there was perennial snowpack in the uplands throughout Late Cretaceous–early Paleocene time (Dettman and Lohmann, 2000; Fan and Dettman, 2009; Foreman et al., 2011).

Long-term eustatic sea level was relatively constant through the Late Cretaceous and Paleocene (~50 m higher than today; Miller et al., 2005; Van Sickel et al., 2004), and there is evidence for several large, high-frequency glacioeustatic changes throughout the Late Cretaceous (Miller et al., 2003). The last evidence of nearby marginal-marine deposits (i.e., the Medicine Bow Formation–Ferris Formation contact) crops out ~1 km below the base of the study area. During deposition of the Ferris Formation, the Western Interior Seaway retreated over 1000 km to the southeast (Lillegraven and Ostresh, 1990); and, as such, sea level was unlikely to have had a significant impact this far upstream, especially on the deposition of the upper portions of the Ferris Formation.

The Ferris Formation was deposited in a rapidly subsiding Laramide basin, so tectonic movements were certainly active throughout Ferris deposition. Climate in the Hanna Basin during this time is relatively unconstrained regionally, but isotope records suggest that there were several global shifts in temperature during the late Maastrichtian and early Paleocene (e.g., Zachos et al., 2001). This suggests that Ferris rivers were likely subject to changes in either climate or tectonic conditions, or both, during the time period over which clustered sand-body stratigraphy was deposited.
The Ferris Formation in the study area consists of coarse-grained channel-belt sandstones and conglomerates separated by intervals of fine-grained (silt and clay) and carbonaceous overbank floodplain material. The channel deposits represent relatively small stream channels draining coarse sands and gravels from nearby (within 10–20 km) tectonic uplifts to the north and northwest. These streams likely flowed through relatively wet floodplains, as evidenced by common dark, organic-rich laminated mudstone deposits, and a lack of bioturbation (Hasiotis and Honey, 2000; Retallack, 2001). Given the coarseness of channel fills, the overbank deposits in the successions are noticeably fine-grained, even close to adjacent channel-belt deposits. Heterolithic deposits that form during progradational-style channel avulsions (e.g., Kraus and Wells, 1999; Slingerland and Smith, 2004; Smith et al., 1989) are notably absent in the Ferris Formation, and sandy crevasse splay units are rare in the successions (Jones and Hajek, 2007). Coarse-grained fluvial sand bodies in the study area record the position of each river channel belt as it migrated and avulsed across the floodplain.

A relatively small portion of the basin containing over 100 individual fluvial sand bodies was chosen for this study. Because of the steep regional dip, the distribution of sand bodies can be seen in aerial photographs (Fig. 2) and is statistically clustered (Hajek et al., 2010). This means that channel deposits are distributed such that over relatively short distances (hundreds of meters), there are many more channels than would be encountered if the same number of sand bodies were distributed randomly across the same area. The primary focus of this study is to elicit the controlling factors responsible for producing this distinctive, clustered stratigraphic pattern.

**DETERMINING CONTROLS ON FERRIS FORMATION ALLUVIAL ARCHITECTURE**

Previously, Hajek et al. (2010) used a statistical framework to describe the distribution of sand bodies within the study area. Here, the goal is to determine whether sedimentologic characteristics of channel-belt sand bodies change systematically in relation to stratigraphic position in the basin. If alloogenic factors controlled the distribution of channels during Ferris deposition, we would expect sedimentologic properties of channel deposits to vary in spatial/stratigraphic relationship with the clustered pattern of sand bodies. In contrast, if Ferris sand-body architecture is a product of autogenic organization, channel properties may have no spatial relationship with the observed stratigraphy.

**Expected River Responses to Changing Boundary Conditions**

In order to conduct this type of stratigraphic evaluation, a set of channel characteristics that would potentially be affected by changing basin boundary conditions must be selected. To identify these key characteristics, we considered the potential stratigraphic and sedimentologic ways in which rivers have been proposed to change in response to allogenic forcing, including climate, tectonics, and sea level. Ultimately, changes in external controls acting on upstream river catchments can impact source area relief, composition, sediment flux, and/or water discharge. In contrast, changes in downstream controls (i.e., base level) are likely to impact the geometry of depositional sequences, erosional surfaces, and floodplain characteristics by forcing localized erosional/aggradational cycles.

Climate changes can directly affect river networks by altering the amount of water and sediment delivered to watersheds (e.g., Gasparini et al., 2006; Schumum, 1977), and also by changing source-area weathering and floodplain characteristics. Increased water discharge could lead to larger rivers, which form larger sand bodies (e.g., Gibling, 2006) and deposit thicker bar clinoforms (e.g., Mohrig et al., 2000). Increases in water discharge relative to sediment supply could also result in erosion and bypass within fluvial systems, forming incised valleys (e.g., Blum and Aslan, 2006; Demko et al., 2004). Increased stream power could also produce coarser channel deposits, as long as suitable grain-size distributions were available in source catchments. Climatic changes can also affect floodplain characteristics, including soil development and habitats for sediment-churning biota (e.g., Hasiotis and Honey, 2000; Kraus, 2002; Smith et al., 2009). Additionally, a climate-controlled increase in weathering rates, particularly the conversion of feldspars to clay, can cause a shift in composition of sediment by supplying more quartz-rich sand and/or more clays to a river basin. Such a supply change can also influence channel-belt geometry by changing river-bank cohesion. Greater mud content provides increased floodplain cohesion (as can increased vegetation cover), which can promote channel narrowing (e.g., Braudrick et al., 2009; Davies and Gibling, 2011; Edmonds and Slingerland, 2010; Tal and Paola, 2007).

Episodic tectonic uplift of mountain ranges can affect fluvial deposition through imposed slope changes, changes in sediment supply (grain size, amount, and composition) delivered to a basin, and by altering the space for sediment accumulation by changing subsidence rates. For example, tectonic tilting may result in changes in grain size and/or water depth of attendant rivers (e.g., Heller et al., 2003), potentially changing the grain size and bar clinoform thickness within deposited sand bodies. Likewise, drainages could be reset or diverted by fault movements (e.g., Schumm et al., 2000) and manifested as changes in paleocurrent directions and possibly composition in the proximal basin (e.g., Jones et al., 2004). Additionally, uplift resulting in channel incision could produce prominent scour surfaces within a fluvial succession (e.g., Holbrook et al., 2006). Changes in basin subsidence could also affect the aspect ratios of preserved channel-belt sand bodies, particularly when limited accommodation promotes fluvial reworking, potentially generating wider sand bodies (e.g., Allen, 1978; Bridge and Leeder, 1979; Leeder, 1978). Furthermore, subsidence patterns influence the supply of particle sizes downstream because larger clasts are preferentially deposited proximally in alluvial basins (Fedele and Paola, 2007; Paola et al., 1992; Strong et al., 2005).

Sea-level changes can also affect fluvial channel deposits, particularly by controlling accommodation and the development of incised valley cut-and-fill cycles (e.g., Blum and Aslan, 2006; Holbrook et al., 2006). Incised valleys may form relatively large sand bodies (Gibling, 2006) and also can result in enhanced paleosol development on stranded interfluvies (e.g., Wright and Marrott, 1993). Additionally, sea-level fluctuations could result in unconformities within a stratigraphic succession (Holbrook et al., 2006). Changes due to sea level, however, would decay upstream of the shoreline as scaled by a river’s backwater distance (e.g., Burns et al., 1997).

These external controls are manifested in preserved deposits in a variety of sedimentological (e.g., paleoflow depth and paleocurrent direction indicators), compositional (e.g., gravel composition), and stratigraphic (e.g., nature and extent of unconformities) characteristics that are measurable in the field (Table 1). As seen in Table 1, it is difficult to uniquely attribute changes in most channel properties with a specific causal mechanism (for example, a change in grain size could be related to either a climate or tectonic change). Nonetheless, any relationship between these properties and the stratigraphic organization in the Ferris Formation, either within or between clusters, or as a function of overall stratigraphic position, would strongly suggest that some external factor controlled deposition of clustered stratigraphic architecture in the unit.

**Field Observations**

Sand-body locations and geometries in the Ferris Formation were mapped using a differential global positioning system (DGPS; accuracy
<table>
<thead>
<tr>
<th>Field Measure</th>
<th>Observation</th>
<th>Climate implication</th>
<th>Tectonic implication</th>
<th>Sea-level implication</th>
</tr>
</thead>
<tbody>
<tr>
<td>Channel sand body size</td>
<td>Sand body width, thickness, and W/T ratio</td>
<td>Larger sand bodies indicate bigger river systems (more discharge).</td>
<td>Large, amalgamated sand bodies reflect increased reworking during relatively slow subsidence.</td>
<td>Large sand bodies resulting from incised-valley filling.</td>
</tr>
<tr>
<td>Clinoform thickness</td>
<td>Thickness of bar clinoform deposits within sand bodies.</td>
<td>Increase with increasing discharge.</td>
<td>Larger clinoforms in mature, integrated catchments.</td>
<td>General increase in clinoform thickness seaward (in tributary systems).</td>
</tr>
<tr>
<td>Paleo current direction</td>
<td>Paleo flow azimuth indicated by trough cross-bedding.</td>
<td>No clear relationship.</td>
<td>Changes with source area or local damming of sediments.</td>
<td>No clear relationship.</td>
</tr>
<tr>
<td>Maximum clast size</td>
<td>Average 6 axis of five largest clasts measured within a sand body.</td>
<td>Increase with increasing discharge.</td>
<td>Increase with pulses of tectonic uplift and increased supply; increase in downstream-fining rate with subsidence increase.</td>
<td>Generally decrease downstream; decreases with rising sea level.</td>
</tr>
<tr>
<td>Unconformities</td>
<td>Erosional surfaces extending beyond the base of individual sand bodies.</td>
<td>Indicate intervals of increased water discharge relative to sediment discharge.</td>
<td>Formed by fluvial incision caused by uplift.</td>
<td>Channel incision and downcutting caused by relative sea-level fall.</td>
</tr>
<tr>
<td>Paleosols</td>
<td>Changes in clay abundance, color, or texture of overbank deposits.</td>
<td>Reflect floodplain composition, wetness, and basin precipitation patterns.</td>
<td>Influenced by subsidence-controlled aggradation rates; high aggradation rates produce weak paleosols.</td>
<td>Relative sea-level fall promotes strong soil development along interfluves.</td>
</tr>
</tbody>
</table>

Note: Summary of key sand-body properties (first column) with the potential to record changes in basin boundary conditions, including climate, tectonics, and sea level. The second column contains brief descriptions of the field observations or measurements required for each property in the Ferris Formation. The remaining columns describe how each property might reflect changing climate, tectonic, or sea-level conditions, respectively, within the basin.
not be confidently identified). Sand-body W/T varies from 3.2 to 104 over the study area, with a mean of 32.1 (n = 120; Fig. 4A).

**Mean Bar Clinoform Thickness**

In active rivers, bars grow to the water surface during channel-forming or bankfull discharge; consequently, preserved bar clinoform deposits are excellent indicators of local paleoflow depth (e.g., Bhattacharya and Tye, 2004; Davidson and North, 2009; Mohrig et al., 2000; Paola and Mohrig, 1996). Bar clinoforms were identified as inclined bedsets within sand bodies, often internally containing multiple sets of cross-beds with the same orientation (e.g., Hajek and Heller, 2012; Mohrig et al., 2000). Clinoforms that show updip rollover, or flattening, and grade upward into finer-grained deposits are fully preserved and can be used to estimate paleoflow depth. The thicknesses (vertical relief) of fully preserved bar clinoforms were measured throughout the study area (Fig. 5). Relief of truncated clinoforms was also measured, providing a minimum estimate of local maximum (bankfull) flow depths. Due to the nature of exposures in the field area, errors on bar clinoform thickness measurements are conservatively presumed to be ±10 cm.

In total, we collected 113 clinoform measurements throughout the study area from 47 different sand bodies. In the remaining sand bodies, clinoform deposits meeting the criteria were not found. Multiple measurements from a single sand body were averaged (number of measurements per sand body ranged from 1 to 8). Mean bar clinoform relief throughout the study interval was 0.59 m, indicating that rivers were relatively shallow throughout the Ferris succession studied (Fig. 4B).

**Mean Paleocurrent Direction**

Paleocurrents were measured primarily from cross-bed features exposed within channel sand bodies (Fig. 5), with most measurements taken from trough cross-beds found in the middle of, or near the base of the channels. Because of the steep dip of the sand bodies, the rake of the trough direction was recorded and restored to horizontal by rotating out the average strike and dip of the sand bodies (084°, 80°S). In total, 140 paleocurrent measurements were collected throughout the study area taken from 45 individual sand bodies.
bodies. The remaining sand bodies lacked well-exposed paleocurrent indicators. Mean paleocurrent azimuth is 146° (Fig. 4C).

**Gravel Clast Size**

Maximum clast size can be a sensitive indicator of fluvial competence and paleo–river slope, which may be related to paleodischarge and paleo–sediment supply conditions (e.g., Hajek and Wolinsky, 2012; Heller et al., 2003; Paola and Mohrig, 1996). In the Ferris Formation, maximum clast size was measured in sand bodies containing substantial pebble fractions (e.g., Fig. 6), and sand/granule size was recorded for sand bodies lacking abundant pebble clasts.

Throughout the Ferris Formation study area, b-axes of the largest gravel clasts (pebbles and larger) were recorded in well-exposed patches at the base of sand bodies (Fig. 6). If the grains were not accessible in three dimensions, the two exposed axes were measured and averaged. In order to characterize the greatest range in maximum clast size across the study area, sizes reported here are the average of the five largest measurements in each sand body. In sand bodies lacking sizable patches of grains larger than several millimeters, maximum sand or granule size was recorded.

Maximum gravel measurements were obtained for 37 of the sand bodies distributed throughout the study area and ranged from 4.0 mm to 73 mm. Maximum particle size of the remaining 87 channels was dominantly 1–3 mm (Fig. 4C).

**Gravel Composition**

In order to identify potential changes in sediment supply source area, sediment composition was measured throughout the study area. Building on provenance work in the Ferris Formation by Ryan (1977), and following the methods of Jones et al. (2004), compositions of 100+ coarse (~10 mm and larger) clasts were recorded for channels with substantial gravel content. Gravel-composition surveys were conducted along bedding planes exposed at the base of sand bodies (Fig. 6). Survey grids were 1 m², and clast composition was recorded every ~10 cm in across the grid until at least 100 clasts had been categorized.

Thirteen sand bodies within the Ferris Formation had sufficient gravel content to conduct full composition surveys. Dominant lithologies in the interval are granite, quartz, and feldspar clasts derived from nearby basement-cored Laramide uplifts, and sundry sedimentary clasts, including Pennsylvanian Tensleep Sandstone, distinctive clasts of the lower Upper Cretaceous Mowry Shale, and reworked chert pebbles from the Lower Cretaceous Cloverly Formation.

**Floodplain Characteristics: Erosion Surfaces and Paleosols**

Unconformities and paleosols were observed throughout the study interval. Erosional contacts between sandstones and underlying floodplain deposits were traced during mapping. Paleosol development, color, grain size, sedimentary structures, and bedding continuity were evaluated in floodplain exposures.

**Statistical Analysis**

Measurements for sand-body dimensions, bar clinoform thicknesses, paleocurrent directions, and maximum clast sizes are compared using three different approaches. First, data for each sand body are plotted with respect to the position of the centroid of the sand bodies to which they belong (Fig. 7). This allows for quick visual inspection of any areal trends in the data. In the second approach, data are broken into subgroups of 50-m-thick stratigraphic intervals. This emphasizes overall changes in parameters that take place over time, where stratigraphic order is a proxy for time, regardless of where laterally in the cross section the channels are located (Fig. 8A). Finally, data are subgrouped by individual sand-body clusters (Fig. 9A). Cluster groups were interpreted visually from their mapped position and have scales consistent with stratigraphic organization statistically identified by Hajek et al. (2010) and Wang et al. (2011). Several cluster group arrangements were also evaluated and showed the same results as reported here, which come from the finest cluster division (i.e., the most cluster groups with scales consistent with previously identified statistical organization within the unit). This approach emphasizes the importance of proximity of channel belts as a control on the data. The hypothesis here is that nearby channels are genetically related and share common source area influences that may differ from other clusters.

In order to determine whether there are population differences between stratigraphic and cluster groups, standard analysis of variance (ANOVA) statistical analysis was used to determine if mean values for subgroups were different than the mean value of the entire population (e.g., Verzani, 2005; Tables 2 and 3). If, for a given variable, the probability that any subgroup could have been randomly drawn from the full data set is greater than 0.05, there is no statistical difference between the subgroups at a 95%
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Confidence interval. ANOVA comparison was conducted using the statistical package R. Bar clinoform measurements, sand-body dimensions ($\log(W/T)$), and grain-size measurements in phi scale were approximately normally distributed, so normal distributions were assumed in ANOVA analysis, and a circular-normal distribution was used to evaluate paleocurrent data. The results for each type of data are summarized next, along with comments about sampling issues that may influence interpretations related to the overall hypothesis.

RESULTS

Sand-Body Dimensions

Sand-body aspect ratios show no strong spatial trends in either stratigraphic or cross-basin directions when plotted by centroids (Fig. 7A). Likewise, there are no stratigraphic sand-body W/T trends (Fig. 8B), nor are there significant differences in sand-body dimension between cluster groups (Fig. 9B). ANOVA analysis reveals statistical differences in sand-body dimensions between cluster groups (Table 2); however, this is attributed to one outlier group (group D), which has a particularly low mean (Fig. 9B).

We note that group D is situated in a particularly windy corridor within the field area, which is in one of the windiest areas of a windy state. In this area, modern erosion of sandstone ridges is especially high, and lateral contacts defining true sand-body widths are not readily exposed. The fact that this more heavily weathered group shows smaller (narrower) sand bodies than the others is not surprising, and we consider this an anomaly unrelated to differences in original, deposited sand-body size in the study area.

Paleoflow Depth

Paleoflow depth, as determined by bar clinoform relief, shows a large amount of scatter throughout the succession (Fig. 7B). Paleoflow depth estimates appear to decrease from 0.65 cm in the lower half of the study area to 0.50 cm in the upper half of the study area; however, ANOVA results indicate there is no statistical difference between stratigraphic groups (Fig. 8C). ANOVA results suggest some statistical difference in paleoflow depth between cluster groups (where clusters B and E seem to be outliers with relatively large and small average paleoflow depths, respectively; Table 3).

Note that sample sizes for some of the groups are quite small (Fig. 9C), and also that the potential error in clinoform measurements (estimated to be ±0.10 m) should be considered when interpreting the significance of these results.

Paleocurrent Direction

Paleocurrent azimuth appears to trend from south-southwest in the lower part of the study interval to south-southeast in the upper part of the study interval (Fig. 7C). Two outliers trending west-northwest appear near the bottom and top of the study area and are located near the eastern side of the field area. Paleocurrent trends show subtle shifting through the stratigraphic succession (Fig. 8C; Table 2). Likewise, there are statistical differences in paleocurrent orientation between cluster groups, where clusters B and E are the most different from other clusters (Fig. 9C; Table 3).

Some cluster and stratigraphic groups have relatively few channels for which mean paleocurrent data were available, so sample sizes are small.

Maximum Clast Size and Gravel Composition

Maximum gravel size shows a wide scatter and lacks spatial trends across the study area (Fig. 7D). The remainder of the sand bodies in each cluster or stratigraphic group have maximum clast sizes less than a few millimeters (fine granule), and are dominantly sand sized. Sand-body clast sizes are variable, but generally the full range of grain sizes is represented in each stratigraphic group with no obvious large-scale trends in the succession (Fig. 8D; Table 2). Additionally, grain size does not vary between cluster groups (Fig. 9D; Table 3). Gravel composition shows no secular trends within the study area, and the variability of multiple counts from one sand body is similar to that found throughout the study area.
entire study area (Fig. 10). These provenance results are consistent with previous studies in the area (Ryan, 1977).

Floodplain Characteristics: Erosion Surfaces and Paleosols

Erosional surfaces in the study interval consist of basal channel scours, and no unconformities that extend beyond the width of the overlying channel sandstone were found. Channel-margin deposits were conformable with lateral floodplain deposits, and in the best exposures, beds could be traced from channel-belt sandstones into distal overbank deposits.

Floodplain deposits in the study interval range from dark gray to black and contain abundant carbonaceous debris and occasional thin (<10-cm-thick) coal seams. Floodplain deposits lack well-developed pedogenic indicators or significant bioturbation, and original bedding (including fissile shale deposits) is routinely preserved. Slight variations in floodplain character, including decreasing grain size and increasing carbonaceous material, occur laterally away from channel margins over scales of 10–20 m. Other than these changes, no significant stratigraphic or lateral variability in paleosol character was observed within the study area.

INTERPRETATION OF SPATIAL TRENDS

Neither sand-body dimensions nor channel grain sizes vary systematically in either the stratigraphic intervals or between cluster groups within the study area. The variability observed in these channel properties may reflect the range of natural variation within Ferris rivers over basin-filling time scales. Paleosol characteristics showed no large-scale spatial trends in the basin and indicate that Ferris floodplains were generally wet and poorly drained (e.g., Retallack, 2001). Likewise the overall lack of pedogenic modification within the succession suggests that aggradation rates were sufficiently high to inhibit significant paleosol development on floodplains (e.g., Kraus, 2002). Additionally, erosion surfaces in the study area are limited to minor disconformities at channel bases. This means that channel sand bodies are largely the result of fluvial activity in an overall aggradational setting and were not confined to incised valleys. Together, these results indicate that while the Ferris Formation was deposited, the basin experienced relatively high aggradation rates and no major episodes of incision and degradation.

Generally, channel properties are more similar than different between stratigraphic and cluster
Field test of autogenic control on alluvial stratigraphy

groups, even for those variables that show some differences across the study area. Note that although this analysis is based on a relatively large stratigraphic data set, the number of well-sampled individual sand bodies is limited. Consequently, when data are lumped into stratigraphic or cluster groups, some groups contain relatively few channels with measurements (e.g., only four sand bodies contain paleocurrent measurements in stratigraphic group II and cluster group B). The range of data observed in the best-sampled cluster and stratigraphic groups (e.g., cluster groups C and G) is relatively large. If more data could be obtained for the groups with small sample sizes, the range of observed values from those groups might also increase, potentially eliminating the apparent paleocurrent and paleoflow depth trends observed within the study area. However, based on presently available data, there are statistical differences in paleocurrent direction and paleoflow depth within the Ferris Formation. Next, we discuss interpretations of stratigraphic trends and differences between cluster groups.

**Stratigraphic Trends**

Paleoflow depth remained the same throughout Ferris deposition. This may not be surprising given that measured paleoflow depths only vary by a factor of two within the study interval. This relatively low degree of variability might reflect the depositional setting, where presumably many small channels draining active uplifts formed a fluvial fan (cf. Hartley et al., 2010; Weissmann et al., 2010). Paleocurrent direction is generally southward within the study area; however, there is some suggestion that rivers shifted subtly through time, from south-southwest to south-southeast and back, as the basin filled. This observed pattern may indicate compensational basin filling, where depositional systems move across a basin to fill accommodation (e.g., Straub et al., 2009; Wang et al., 2011).

**Cluster Group Trends**

Both paleocurrent and paleoflow depth data show some difference between cluster groups, suggesting that different clusters may have been deposited by different individual river systems.
There are differences in paleocurrent direction between some of the cluster groups, and even between clusters that occupy the same stratigraphic intervals. For example, clusters A and B comprise stratigraphic group I at the base of the studied succession (Figs. 8A and 9A), yet they show different paleocurrent measurements, with A originating dominantly from the north-northwest and B originating from the northeast (Fig. 9D). Additionally, these clusters have slightly different average paleoflow depths, with measurements from B (mean = 0.71 m) being higher than those from A (mean = 0.56). This result indicates that some clusters may have formed contemporaneously by separate fluvial systems originating from somewhat different drainage areas along the same mountain front. Therefore, the study area may represent a cross-section through coalescing fluvial fans filling the proximal basin. This interpretation cannot be confirmed with composition data, which does not vary significantly within the study interval.

**DISCUSSION**

The Ferris Formation in the study area is characterized by well-developed clustering of channel-belt sand bodies. However, in general, the variability of observed channel properties within the Ferris Formation is relatively large, and channel properties are, at most, weakly related to the alluvial architecture pattern observed in the basin. Consequently, there is no strong evidence of direct extrinsic control on channel clustering in the Ferris Formation. As such, there is no need to evoke external forcing as the cause of clustering within the Ferris Formation. This is not to say that climate, tectonics, or sea level did not change throughout Ferris Formation deposition. Rather, stratigraphic and sedimentologic signals of changing basin boundary conditions were overwhelmed by intrinsic variability within the Ferris depositional system.

This follows the concept of “signal shredding” (Jerolmack and Paola, 2010), whereby a critical autogenic time scale exists, below which most external signals are obliterated by autogenic dynamics within the sedimentary system. The only signals that can pass through the autogenic filter and become preserved in the stratigraphic record are those with wavelengths exceeding the critical autogenic time scale or large events that far exceed the amplitude of the maximum autogenic fluctuations possible within a given system. Channel avulsion is a primary source of autogenic variability in
TABLE 2. ONE-WAY ANOVA RESULTS FOR STRATIGRAPHIC GROUPS

<table>
<thead>
<tr>
<th></th>
<th>Degrees of freedom</th>
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<th>Mean square</th>
<th>F value</th>
<th>Pr(&gt;F)</th>
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<tbody>
<tr>
<td>Sand-body width/thickness</td>
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Mean bar clinoform thickness

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Mean paleocurrent azimuth

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Maximum clast size

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Note: One-way analysis of variance (ANOVA) results for field measurements in stratigraphic groups. Degrees of freedom are defined as number of groups minus 1 for groups, and the total number of observations for the whole population. Pr(>F) indicates the probability that any individual group represents a random sample of the overall population. A probability of less than or equal to 0.05 is considered significant in this case (analogous to a 95% confidence interval). Note that a circular ANOVA test was used for azimuthal paleocurrent data. Comments indicate whether or not there is a statistical difference between groups and any sampling issues.

TABLE 3. ONE-WAY ANOVA RESULTS FOR CLUSTER GROUPS

<table>
<thead>
<tr>
<th></th>
<th>Degrees of freedom</th>
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<th>Mean square</th>
<th>F value</th>
<th>Pr(&gt;F)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sand-body width/thickness</td>
<td>6</td>
<td>2.08</td>
<td>0.35</td>
<td>2.76</td>
<td>0.02</td>
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Mean bar clinoform thickness

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<th>F value</th>
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Mean paleocurrent azimuth

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Maximum clast size

<table>
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<th>Sum of squares</th>
<th>Mean square</th>
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<th>Pr(&gt;F)</th>
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Note: One-way analysis of variance (ANOVA) results for field measurements in cluster groups. See note in Table 2 for further details.
Figure 10. Gravel composition. Locations of gravel surveys are indicated in upper map. Bar graphs indicate proportion of clasts in each gravel count. Numbers below bar graphs indicate sand body. Multiple surveys were conducted in different locations on some sand bodies (e.g., #2). Granite includes granite clasts as well as quartz and feldspar particles. Chert includes reworked chert pebbles from the Cloverly conglomerate, and “Mowry” indicates distinctive clasts re-worked from the Mowry Shale. “Other” indicates other, dominantly sedimentary clasts, including reworked particles of Tensleep Sandstone.

References Cited


