ORIGINS OF BIMODAL STRATIGRAPHY IN FLUVIAL DEPOSITS: AN EXAMPLE FROM THE MORRISON FORMATION (UPPER JURASSIC), WESTERN U.S.A.

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ABSTRACT: Changes in stacking density of fluvial channel-belt sand bodies are one of the most obvious aspects of alluvial architecture. The proximity of sand bodies to each other has been used as a field criterion to suggest major changes in external factors, such as basin subsidence rates, relative changes in base level, and changes in climate. In places rapid changes in stacking density are found in vertical succession between very high to very low net sandstone. In the absence of an intervening unconformity such changes are not the result of two unrelated, but juxtaposed, fluvial regimes but represent a more transitional change of formative conditions or preservation style.

To study whether a rapid change in stacking density was controlled by external factors rather than autogenic behavior, we sought to document other possible changes in deposition that took place in concert with changes in stacking pattern. The Morrison Formation, Upper Jurassic of eastern Utah and western Colorado, U.S.A., contains a rapid transition from the Salt Wash Member, with an average 63% net sand, to the overlying Brushy Basin Member, dominated by muddy floodplain deposits (10% net sand). We measured nine sedimentary characteristics related to fluvial system properties in both members along a down-paleoflow-oriented transect. The only significant change found between the two members is the composition and abundance of floodplain clay minerals. The transition in stacking density is coincident with a rapid increase in smectitic clay, interpreted to result from increased volcanic ash influx, at the expense of illitic clay. The ash was preferentially accumulated in the floodplain, either as airfall or carried in the wash load of rivers.

Geometric calculations suggest that in order to generate the observed change in stacking pattern by the addition of volcanic ash alone, more than half of the floodplain deposition would have to be composed of ash. An increase in ash supply would increase floodplain sedimentation rate and reduce the frequency of avulsion. Together these effects could reduce the stacking density by the requisite 50%.

Dependence of stacking density on floodplain sedimentation rate highlights the danger in assuming changes in stacking density observed in alluvial sequences are mainly controlled by other external factors such as sea-level fluctuation, subsidence rate, or overall aggradation rates.

INTRODUCTION

Alluvial systems construct channel boundaries out of their transported sediment loads. Grossly simplified, these form two distinct depositional environments—channel belts and floodplains. Floodplains are chiefly composed of fine-grained sediment (mud) carried out of a channel in suspension. Channel-belt sediment includes sand and gravel transported intermittently in tractive and suspended modes. Fluvial stratigraphy is composed of a mosaic of these depositional building blocks. Sequences relatively abundant in coarse channel-belt deposits are said to have high stacking density or high net sand. Those with relatively abundant floodplain deposits have a low stacking density or low net sand.

Some fluvial stratigraphy exhibits a striking change in stacking density where there is a rapid vertical transition from very sandy deposits composed of channel-belt sandstones to very muddy, floodplain-dominated deposits. Two examples from the Rocky Mountains of the western U.S.A. include the Morrison Formation (Late Jurassic) and the Wasatch Formation (Paleogene) (Fig. 1). Both examples show rapid transition between mudstone-rich, floodplain deposits (c. 10% net sand) to very sandstone-dominant, channel-belt deposits (c. 65% net sand) in vertical sequence with no intervening unconformity.

Various models have been proposed to explain rapid changes in stacking density or bimodal stratigraphy, including changes in basin aggradation (subsidence) rate, climate-induced changes in water supply, sediment supply and vegetation along extant river systems, and changes in base level (Allen 1978; Leeder 1978; Bridge and Leeder 1979; Olsen et al. 1995; Heller and Paola 1996; Demko et al. 2004; Gibling 2006; Kjemperud et al. 2008; Davies and Gibling 2010). Alternatively such abrupt changes may be a form of self-organization caused...
The purpose of this study is to evaluate the possible cause of bimodal stratigraphy seen in one clear example—the Morrison Formation (Upper Jurassic) in east-central Utah and adjacent parts of Colorado. Is the observed change in stacking pattern coincident with other changes that reflect synchronous external control on deposition? Alternatively perhaps gradual changes in the depositional system caused an abrupt change in stacking density as some threshold was crossed. If there are no changes in other aspects of sedimentation, an autogenic cause for change in stacking density can be inferred.

Our approach is to identify those changes in depositional character that take place in concert with changes in stacking pattern. If such changes exist and vary monotonically with stacking pattern, we assume that there is external control and that the deposits record these changes. However, if changes in stacking density are uncoupled from changes in other aspects of deposition, a different mechanism must be invoked. To the degree that changes in stratigraphic character are used to infer changes in boundary conditions, it is important to evaluate the validity of the connection.

**BIMODAL STACKING PATTERNS**

Changes in channel-belt stacking density are one of the most obvious aspects of fluvial successions (Fig. 1) and much has been written on the possible causes of these changes (Allen 1978; Leeder 1978; Bridge and Leeder 1979; Wright and Marriott 1993; Shanley and McCabe 1994; Foreman et al. 2012). When analyzing stacking densities of fluvial units in the central Rocky Mountains we find that, while the entire range of stacking densities occur (Appendix), most values (58%) fall either between 0–20% and 80–100% net sand. This slight bimodality is enhanced by the observation that in some cases stacking density changes rapidly from very mud-rich to very sandy in vertical succession, such as seen in Figure 1.

Where bimodality in stacking density occurs, the rapid transition suggests three different possible controls on this behavior. (1) A sudden change in some external controlling variable such as climate or tectonics took place that generated a synchronous change in the stratigraphic record. In order for this change to form a rapid transition in stacking density, the time scale of change must be short relative to the stratigraphic response time (Paola et al. 1992). In this case, the observed change in stacking pattern is coincident with the change in boundary conditions. (2) Alternatively, the change in stacking density was a threshold response to a more gradually changing external control. Here the change in stacking density reflects a critical value that, once passed, generated a new stacking pattern. In this case, the stratigraphic change is not coincident with the onset of a new external control, but reflects a rate of change in magnitude of that control. (3) Lastly, the change in stacking density may have been entirely uncoupled from external controls and, instead, resulted from autogenic processes within the sedimentary transport system. For example, Hajek et al. (2012) describe scales of self-organization of river avulsion, wherein clusters of channel belts can locally form high stacking densities without changes in boundary conditions.

In order to evaluate these possible controls on channel stacking patterns, we focused on alluvial architecture of the Morrison Formation in part of the U.S. Rocky Mountains. The unit was chosen because it is well exposed over a broad region and exhibits bimodal stacking density (Fig. 2), which averages 63% net sandstone in the Salt Wash Member and about 10% net sandstone in the Brushy Basin Member. Our approach was to observe what types of other changes took place in the depositional system in concert with the change in stacking pattern.

**MORRISON FORMATION**

The unit of focus for this study is the Morrison Formation of Late Jurassic age (Kimmeridgian–Tithonian; Turner and Peterson 2004) exposed in eastern Utah and west-central Colorado (Fig. 3). In this area the formation contains bimodal stratigraphy that is regionally extensive and easily accessible. The unit was deposited mostly by rivers emanating from the evolving continental volcanic arc and associated thrusted terranes to the west and the Mogollon Highlands to the south of the depositional basin (Yingling and Heller 1992; Turner and Peterson 2004; Dickinson and Gehrels 2008; Fuentes et al. 2009). In the study area, data on paleocurrents (Fig. 3) and detrital provenance

![Image](54x235)
(Dickinson and Gehrels 2008) indicate sediment derivation from the west-southwest.

In eastern Utah, the Morrison Formation overlies a subtle unconformity (J-5; Pipirinos and O’Sullivan 1978), below which playa deposits of the Summerville Formation represent the final marine regression towards the northwest (Blakey et al. 1983). The basal Tidwell Member of the Morrison Formation is a thin unit that includes lacustrine, evaporite, mudflat, with minor alluvial and eolian deposits (Petersen 1988) and, towards the north, tidal deposits, that mark the progradation of a fluvial system across marginal marine environments (Turner and Peterson 2004). The Tidwell Member is overlain by the Salt Wash Member, a sand-rich fluvial unit characterized by abundant channel-belt sand bodies, with an average stacking density of 63% (Fig. 2). The unit thickens rapidly towards its most upstream exposure, towards the southwest (Fig. 4A), and has been interpreted to represent a fluvial megafan (Owen et al. 2015). Available age data indicate that the Salt Wash reached its maximum extent during middle Kimmeridgian time (Turner and Peterson 2004).

Sitting on top of the Salt Wash Member is the Brushy Basin Member of the Morrison Formation (late Kimmeridgian–Tithonian; Turner and Peterson 2004). The Brushy Basin Member is notably finer-grained than the underlying Salt Wash Member (Fig. 2) and is dominated by overbank siltstones with paleosols and finer-grained flood-plain lakes, which become more common upstream (Turner and Fishman 2001; Demko et al. 2004).

Paleosols are found throughout the Morrison Formation but are especially well developed in the Brushy Basin Member (Demko et al. 2004). One well-developed paleosol is found in many parts of the study area at or near the boundary between the Salt Wash and Brushy Basin members. Demko et al. (2004) interpreted this paleosol as marking a regional unconformity, the “Mid-Morrison Unconformity,” however Kjemperud et al. (2008), as well as our own observations, noted places where channel belts and groups of channel belts similar to those seen in the Salt Wash Member either cut through or are found within the paleosol horizon, suggesting that it did not record a major episode of regional erosion.

Isopach patterns show that the Brushy Basin Member thins to the southwest (Fig. 4B) and is generally thinner where the underlying Salt Wash Member is thickest. The isopach pattern of this member may have been impacted by a regional unconformity identified at the top of the Morrison Formation across the region, separating the unit from overlying Lower Cretaceous deposits (Demko et al. 2004).

In the study area, the ages of the Salt Wash and Brushy Basin members are constrained by 40Ar/39Ar sanidine and U-Pb zircon ages from airfall or reworked ash beds. These ages range from 156 ± 4 Ma to 148 ± 2 Ma (Bradshaw and Kowallis 2009 2010). Recalibrated 40Ar/39Ar ages and regional correlation (Trujillo and Kowallis 2015) indicate that sedimentation rates in the Brushy Basin Member in the study area were c. 57 cm/ka, about five times higher than those of the Salt Wash Member (c. 12 cm/ka).

Most studies suggest that the climate became progressively wetter throughout deposition of the Morrison Formation (Demko et al. 2004; Turner and Peterson 2004; Myers et al. 2014). Demko et al. (2004) found argillic calcisols in the Salt Wash and lower Brushy Basin members that range in thickness and maturity. These paleosols become more weakly developed in the upper Brushy Basin Member, indicating a wetter climate. Also, the eastern extent of the Brushy Basin Member contains wetland soils, indicating that more water was available during deposition (Dunagan and Turner 2004). An increase in humidity over time is consistent with the gradual northward drift of North America so that this region moved from subtropical arid conditions into the mid-latitude belt of westerly winds (Dickinson and Gehrels 2008).

A marked increase in smectitic clay and/or smectite-rich mixed-layer clay in the upper part of the Brushy Basin Member has been identified, coincident with a decrease in illitic clay (Owen et al. 1989). This clay change was once considered to be a regional stratigraphic horizon in the Morrison Formation; however, more recent studies suggest that the clay mineralogy in the Morrison Formation is more variable (Trujillo 2006). The origin of smectite has been interpreted to result from an increase in input of volcanic ash from eruptions farther west (Turner and Peterson 2004).

**DATA AND ANALYSIS**

We hypothesize that a change in fluvial system properties exists coincident with the observed vertical change in stacking density. In other words, changes in external controls on the alluvial system (sediment supply, hydrologic regime, etc.) will manifest not only in the stacking density but also in the properties of the preserved channel bodies on scales ranging from grain size to bar forms. These hypothesized changes may be either sudden and coincident with the change in stacking density, or in the case of a critical threshold event, more gradual and begin prior to the change in stacking. The types of changes expected are those seen in the river channel fills, such as reconstructed paleoflow depth and modal grain size—channel belts, such as total sand-body thickness, number of stories, sand body width and sinuosity—and in the flood plain, including variations in clay mineralogy and the relative abundance of crevasse-splay deposits.

Data were collected from both the Salt Wash and Brushy Basin members at outcrops found within 50 km of a SW-to-NE-oriented transect that runs from Tropic, Utah, to Grand Junction, Colorado (Fig. 3). This transect is oriented close to average regional paleoflow directions in the Salt Wash and lower Brushy Basin members. Where appropriate, data results are projected onto this line of transect to show upstream to downstream trends. In addition, clay mineralogy was analyzed along three vertical sections (Fig. 3) as described below.
Paleoflow Depth

Channel bankfull-flow depth is affected by changes in precipitation or catchment size as well as changes in river planform (i.e., meandering vs. braiding). A proxy for paleoflow depth is relief of fully preserved bar clinoforms which are commonly found in preserved channel fills (Allen 1965; Willis 1993; Mohrig et al. 2000; Hajek and Wolinsky 2012). While there may be some slight adjustment for compaction, the measurement of relief on fully preserved sigmoidal bar accretion sets (those that include roll over at the bar top) should be close to true local flow depth because the nature of bars is to grow nearly to the river free surface.

Our results are shown in Figure 5A. There is no statistical difference in flow depths between the two units (Kolmogorov-Smirnov test p-value = 0.3315). As well, there is no change in flow depth with distance down basin (slope of best fit lines are nearly horizontal (≤ 10⁻³) for both units). Local variations in bar relief vary in many places by a factor of two.

Grain Size

Among the ways that fluvial systems can adjust to changes in external conditions is by changing the grain-size distribution of material delivered to the basin as well as changing the rate of downstream fining across the basin.
FIG. 5.—Measured A) bar clinoform relief, B) maximum clast size (average of 10 largest clasts), and C) total thickness of individual sand bodies for the Salt Wash and Brushy Basin members of the Morrison Formation projected onto the line of transect (Fig. 3). Best-fit linear regression is shown for grain size for Salt Wash (dashed) and Brushy Basin (solid) members.

FIG. 6.—A) Number of stories per sand body, B) maximum preserved channel-belt width, and C) number of splays per channel-belt sand body for the Salt Wash and Brushy Basin members of the Morrison Formation along transect.
assuming that a range of grain sizes of source material is available (Paola et al. 1992; Hajek et al. 2012). To characterize trends in grain-size distribution within and between the members, we measured and then averaged the intermediate diameter of the ten coarsest grains found in accessible parts of individual stories of channel-belt deposits. This measurement is relatively easy to make, as compared to characterizing the entire grain-size distribution. These grains are assumed to reflect the coarse tail of the material transported as bed load by the formative streams.

Our results show no difference in maximum grain size between the two members (Fig. 5B). Both members show coarse-tail downstream fining grain-size trend of 0.06 ± 0.01 mm/km. Slight downstream fining also has been documented for the Salt Wash Member in a separate study (Owen et al. 2015).

**Total Sandstone-Body Thickness and Number of Stories**

Sandstone bodies in the Morrison Formation are the manifestation of channel belts. Where sandstone bodies are multistory, each story is scaled by river flow depth (Friend et al. 1979) and is assumed to record an individual channel belt which has been amalgamated with other belts to form an individual sand body (Mohrig et al. 2000). Total sandstone-body thickness relates to a combination of factors including aggradation rate, river flow depth, number of stories per sand body, and relief of the channel belt above the nearby flood plain prior to avulsion (Mohrig et al. 2000). Our results (Fig. 5C) show that there is a large range in maximum sandbody thickness by nearly an order of magnitude for both the Salt Wash and Brushy Basin Members. There is no meaningful downstream change in thickness ($r^2 = 0.005$) and no difference between the distributions between members (Kolmogorov-Smirnov test p-value = 0.239).

Channel-belt sandstone bodies observed are normally composed of more than one story. Stories are often composed of completely preserved, or nearly so, bar clinoform deposits. However, in places there are only lower parts of barforms preserved with truncated upper parts. Such truncations reflect migration of channels that bevel preexisting barforms. Counting stories is sometimes difficult to do if subsequent stories are laterally adjacent rather than vertically amalgamated, as seen to occur frequently in the Salt Wash Member. Our results (Fig. 6A) show that most bodies are composed of two-story sandstones, regardless of whether they are in the Salt Wash or Brushy Basin Member. There is no meaningful difference in distribution between the two members (Pearson’s $\chi^2$ p-value = 0.6572).

**Channel-Belt Width and Sinuosity**

Planform exposures of channel-belt sand bodies, such as seen in Figure 7, are found in some parts of the study area. While some authors consider such narrow, sinuous bodies to be individual channels (Owen et al. 2015), we think that such bodies are preserved channel belts (Mohrig et al. 2000). One reason we make this interpretation is that in places where abandoned channel forms (such as mud plugs) are found on these features they are of smaller scale than the entire sandstone body. Secondly, barforms preserved in these bodies generally suggest that channels (whether braided or meandering) migrated. This is not to say that channels and channel belts cannot be of the same scale. Systems that avulse frequently, such as the Assiniboine River in western Canada (Rannie 1990), can preserve abandoned channels that show little evidence of growth through time. However, one might expect abandoned channels to be filled in with finer-grained deposits than those associated with active migration of bars and coarse-grained bedforms. The features we identify contain these coarse-grained deposits, suggesting deposition by active channel flow. We thus assume that the channel did migrate, at least to some extent, within the belts observed.

The preserved belts permit us to determine the preserved maximum width of channel belts, as measured normal to the medial axis. These data show that the sand bodies in both units have a range of width that encompasses an order of magnitude (Fig. 6B). While average widths for both members are similar (37 m v. 43 m), statistically we can mildly reject the hypothesis that widths from both the Salt Wash and Brushy Basin Members came from the same distribution (Kolmogorov-Smirnov p-value = 0.05219).

Previous authors have interpreted the river planform for sand bodies in both members of the Morrison Formation in the study area and elsewhere. Typically

![Figure 7](image_url) — Example of sinuous channel-belt sand body in the Salt Wash Member seen in plan view on Google Earth centered at 38°24'16"N, 111°00'58"W. Paleoflow was from left to right across the image.

![Figure 8](image_url) — Sinuosity of channel-belt sand bodies determined from planform exposure of channel-belt segments. Exposed channel-belt length is measured along the mid-line of sand bodies as seen on Google Earth images.
the Salt Wash Member is interpreted to be deposits from braided streams (Peterson 1984; Robinson and McCabe 1997; Kjemperud et al. 2008), although Hartley et al. (2015) identified sinuous, meandering channel belts in places. The Brushy Basin Member has been interpreted as recording deposition in a variety of river types from meandering, braided to anastomosed (Yingling and Heller 1992; Galli 2003; Demko et al. 2004). Given the great difficulty in interpreting river planform from stratigraphic cross sections (Bridge 1985), we feel uncomfortable in making such assignments. However, we can ascertain the sinuosity of channel belts in both members from channel-belt segments exposed in planform views in Google Earth images (Fig. 7). We note that sinuosity of channel belts is not equivalent to sinuosity of the individual channels that formed the channel belt. Presumably the sinuosity of the individual channels that constituted the channel belt was equal to, or greater than, that of the channel belts themselves.

Segment lengths run from 68 to 3413 m, with those in the Salt Wash Member being shorter, mostly because more segments are cut off or overlapped by younger channel belts in this more amalgamated member. Channel-belts in both units are fairly straight to “twisty” (Leopold et al. 1964), with sinuosity values mostly less than 1.5 (Fig. 8). Low sinuosity values are the rule for short channel-belt segments, where the wavelength of the sinuous path is long relative to exposed segment length. Thus as path lengths become increasingly small, sinuosity becomes increasingly straight. With increasing length, we would expect to see a leveling off of sinuosity values, as seen in the data from the Brushy Basin member (Fig. 8). The overlap between the Salt Wash and Brushy Basin members indicates that channel-belt patterns did not change significantly between the two units (Kolmogorov-Smirnov test p-value = 0.6134).

**Abundance of Splays**

Both the Salt Wash and Brushy Basin members contain crevasse-splay deposits that either sit directly below channel belts or are otherwise dispersed among the flood-plain deposits (Fig. 9). Splay intervals are recognized as being thin, planar, laterally extensive event beds. Splay deposits are found in groups of a few beds or as isolated, individual beds. Crevasse-splay deposits are found in all stratigraphic sections studied. In the case of the Salt Wash Member, 75% of sand bodies are directly underlain by thin intervals of flood-plain deposits containing splay deposits. In the Brushy Basin Member, splay deposits are found directly below channel-belt sand bodies in about 25% of the cases. Otherwise splay beds are found elsewhere in floodplain intervals.

We noted that although the Brushy Basin Member was overall much more mud-rich than the Salt Wash Member, it appeared that the number of splays

FIG. 9.—Photos of sandbody and crevasse-splay deposits. A) Brushy Basin Member exposed along the east side of the San Rafael Swell (38°55.611′ N, 110°22.632′ W). Two channel-belt sand bodies are exposed at the top of the section. Sand body at the top is 5.5 m thick. B) Salt Wash Member exposed at Yellowcat, Utah (38°50.164′ N, 109°32.250′ W) showing parts of three channel-belt sandstone bodies. Note person, bottom center, for scale. In both photos particularly splay-rich intervals are highlighted with double arrows.
observed in stratigraphic sections, per channel-belt sand body, was similar between the two members. In an attempt to quantify this relationship, we counted the number of splay beds observed in large outcrops, ideally traversing the whole vertical section of the member, and the number of discrete channel-belt sand bodies found in the same exposure and present these results as a ratio (Fig. 6C).

This method contains uncertainty in that recognizing every discrete splay deposit can be difficult and the results show a large range of variability (Fig. 6C). Nonetheless, the results are similar for both members, although the Brushy Basin Member has more sections with somewhat fewer crevasse-splay beds.

Clay Mineralogy

Three long cores drilled through the Morrison Formation in the study area were made available by ExxonMobil Upstream Research Company for description and sampling for clay mineralogy. The cores were collected near Shitamaring Canyon (aka Shootaring Canyon), south of Hanksville, Utah (37°45.687'N, 110°42.081'W); west of Green River, Utah (39°1.873’ N, 110°21.948’ W); and south of Grand Junction, Colorado (38°59.073’ N, 108°36.529’ W) (Fig. 2). All of the cores traversed the entire thickness of the Salt Wash Member, and, along with field sampling of the overlying Brushy Basin Member, were described and sampled for clay mineralogy. Samples were collected from the B and C horizons of paleosols in order to isolate those clays associated with formation of the soil horizons versus parent material. Clays were separated using standard techniques including treatment with MgCl\(_2\) in order to saturate the ion-exchange positions with Mg. Samples were then mounted on glass slides and analyzed by X-ray diffraction including both air-dried and glycolated treatment. Smectite 001 peaks were identified and peak height relative to background is used as a relative indicator of abundance of smectite (and/or smectite-rich mixed layer clay) (Fig. 10). The other dominant clay present is illite, which generally decreases in peak height as smectite becomes more abundant upsection (Fig. 10).

These data strongly suggest that the influx of smectite begins coincident with, or slightly earlier than, the boundary between the Salt Wash and Brushy Basin members at all three locations. However, in the Brushy Basin Member the abundance of smectite varies strongly, a result similar to those of other studies of Morrison clay mineralogy (Keller 1962; Owen et al. 1989; Trujillo 2006).

DISCUSSION

Based on the characteristics examined, we conclude that the Brushy Basin and Salt Wash members of the Morrison Formation record equivalent flow depth, grain sizes, channel-belt and story characteristics, and number of preserved crevasse-splay deposits. From these data we infer that there is no discernable difference within the study area between the rivers that deposited the Salt Wash Member and those that deposited the Brushy Basin. At the scale of river channels, the range of flow depth and maximum grain-size trends were the same. Both members show a gradual downstream fining rate but no associated change in paleoflow depth downstream over the 300 km transect. The downstream decrease in grain size coupled with no obvious change in flow depth implies a regional downstream decrease in paleoslope, if equilibrium transport is assumed.

At the channel-scale, density, width, sinuosity, sand-body thickness, and number of stories are the same. There is a mildly significant difference in channel-belt width. This suggests that the rivers that fed these two members were nearly identical in geometry. Therefore the change in stacking density between the members is likely not tied to significant differences in the river transport systems.

In contrast to the channel-belt deposits in the Morrison Formation, there is evidence of a change in aspects of the floodplain deposits in the members studied. One change is that there was a shift towards a wetter climate during Brushy Basin time as seen in paleosol development, increased presence of flood-plain lakes, and ichnofacies (Turner and Fishman 2001; Demko et al. 2004; Hasiotis 2004). However, these changes are found only in the upper part of the Brushy Basin Member, well after the change in stacking pattern had been established. Instead, the most obvious differences are in relative abundance of overbank mudstones between the two members, an increase in paleosol maturity at least in the lower part of the Brushy Basin Member (Demko et al. 2004) and a pronounced increase in smectitic clay. In addition, Trujillo and Kowallis (2015) show that sedimentation rates increased between the time of Salt Wash and Brushy Basin deposition by roughly a factor of five,
Volcanic ash (white).

The Salt Wash Member consists of channel-belt sand bodies (black) and overbank deposits (stipple). The Brushy Basin Member consists of the same abundance of sandbody and overbank deposits, per unit time, as the Salt Wash Member, as well as volcanic ash (white). Although more pronounced paleosol development in the latter member implies that sedimentation became more episodic as well.

**Role of Volcanic Ash**

Clay-mineral analyses show a marked increase in smectite abundance coincident with, or near, the base of the Brushy Basin Member at all three locations studied. While it would be ideal to use these results to calculate relative abundance of different types of clay minerals, there are simply too many variables in clay mineralogy to do bulk composition in any quantitative way, and available techniques are not designed to make such calculations (Yuan and Bish 2010). Nonetheless, the large increase in peak height seen in Figure 10 indicates a marked increase in smectite abundance. This change is inconsistent with simple burial diagenesis of smectite to form illite with depth. Such a change should be gradual with depth of burial and likely take place transitionally, involving a mixed-layer clay transitional state (Hower et al. 1976). The abrupt shift of the smectite peak height instead suggests to us that it results from alteration of influx of volcanic ash into the floodplain sediment, an interpretation consistent with that of Keller (1962) and accords with the locally abundant occurrence of altered ash beds in the Brushy Basin Member (Turner and Fishman 2001; Dunagan and Turner 2004). The increase in abundance of volcanic ash during Morrison deposition is generally consistent with the Middle to Late Jurassic expansion of the volcanic arc in the U.S. Cordillera (Dickinson 2001), although the reason for the relatively sudden increase in ash deposition is unclear.

The abrupt transition into smectite-dominant clays stands in contrast to the observations of Owen et al. (1989) and Turner and Peterson (2004) that the major clay transition is found higher in the Brushy Basin Member. Owen's study, however, was concentrated on deposits from ancient Lake T’oo’di, mostly east of our study area. All studies of clay mineralogy show that smectite, or smectite-rich mixed-layer clay, generally increases up section, although the amounts are highly variable in the Brushy Basin Member (Fig. 10; Keller 1962; Owen et al. 1989; Trujillo 2006).

The marked, rapid increase in smectite abundance coincident with the change in stacking density leads us to consider the role that a large influx of volcanic ash might have on alluvial architecture. We assume that an increase, whether directly from airfall or reworked and transported as wash load by rivers, would lead to an increase in floodplain aggradation rate while having limited impact on channel-belt aggradation rate, which we presume to be dominated by bedload deposition. Increased floodplain deposition rates can have potential impacts on river pattern in several ways. The first possible impact is to increase cohesiveness of floodplain deposits exposed in river banks. While ash itself is not cohesive, rapid alteration to clay by soil diagenesis can significantly increase soil strength (Herrera et al. 2007). Strengthening river banks can promote a change in river planform from braided to meandering (Schumm 1985). However, there is no evidence found in this study to suggest any significant change in channel form between the Salt Wash and Brushy Basin members other than a slight narrowing of channel belt from the former to the latter. Secondly, rapid increase in deposition rate, coupled with a change in soil geochemistry caused by an influx of ash that changes the mineralogy of floodplain materials, documented to have occurred in the Morrison Formation in the study area (Ratigan 2014), could have an impact on type and abundance of vegetation. Vegetation, in turn, can perform two functions: increase root depth and/or matting can increase bank strength, promoting the formation of single, deep channels; and increased vegetative cover can more effectively buffer out-of-channel flow during times of flooding, extracting flow energy and leading to increased sedimentation rate in the proximal floodplain. However, studies of modern volcanic eruptions indicate that plant succession proceeds quickly (del Moral and Lacher 2005; Talbot et al. 2010), suggesting that possible impacts would be, at best, geologic transients in the stratigraphic record.

Lastly, a more compelling case can be made that a significant increase in floodplain aggradation rate could significantly reduce avulsion frequency, translating into a decrease in channel-belt stacking (Heller and Paola 1996). The addition of volcanic ash onto the floodplain yields two effects that reduce stacking density. The first is a simple mass volume effect by which there is more floodplain sediment (yellow) to accumulate between avulsion events.

![Figure 11](image_url)

**Fig. 11.**—Model for calculating amount of ash needed ($\lambda$) to decrease stacking density from 63% (average for Salt Wash Member) to 10% (average of Brushy Basin Member). The Salt Wash Member consists of channel-belt sand bodies (black) and overbank deposits (stipple). The Brushy Basin Member consists of the same abundance of sandbody and overbank deposits, per unit time, as the Salt Wash Member, as well as volcanic ash (white). Although more pronounced paleosol development in the latter member implies that sedimentation became more episodic as well.

![Figure 12](image_url)

**Fig. 12.**—Explanation of how increased aggradation rate of floodplain reduces the frequency of avulsion. A) Minimum critical relief (perching) between the channel belt and floodplain is needed for avulsion to occur. B) If aggradation rate of the channel belt ($a$) is faster than the floodplain ($b$), avulsion occurs after some time interval. In this case, $a = 2b$, and the amount of aggradation (yellow) that occurs while critical relief is achieved is thin. C) If floodplain aggradation is augmented by addition of volcanic ash ($c$), it takes longer for channel belt to reach critical relief. This results in thicker intervals of floodplain sediment (yellow) to accumulate between avulsion events.
calculate how much ash must be deposited to turn the average stacking density of the Salt Wash Member (63%) into the Brushy Basin Member (10%). In this example, 84% of all deposition (or 93% of all floodplain deposition) in the Brushy Basin Member would have to be volcanic ash (either airfall or reworked) in order to create the observed change in stacking density in the absence of other effects. The required increase in average sedimentation rate would need to be 6.3 times to explain the change in stacking density by this effect alone. Trujillo and Kowallis (2015) indicate that the observed change in sedimentation rate between the units is about five times. While certainly not an exact fit, given the simplicity of the geometric approach, the observed values are close enough to model results to suggest that increased influx of floodplain material accounts for most of the change in observed stacking density.

Independently, we can consider the required rate of ash deposition by comparison with recent influx of volcanic ash from active volcanoes. Following the initiation of continental-arc magmatism in Triassic time, the Cordilleran volcanic arc had grown to form a continuous feature along the length of the Cordillera by Late Jurassic time, including widely distributed volcanic centers across what is now eastern Nevada and western Utah (Dickinson 2006). It is not possible to know what the rate of ash supply was in Late Jurassic time; however, we can compare the required rate of ash input with that produced by recent eruptions of ash from Mt. Mazama, Oregon c. 7000 years ago and Mt. St. Helens, on May 18, 1980 (Shipley and Sarna-Wojcicki 1983). These eruptions are volcanic-arc events, similar to what might have occurred in the western U.S. in Late Jurassic time. At distances of about 500 km downwind of these eruptive centers (a distance similar to the palinspastically restored distance from the study area to volcanic centers farther west), about 4.5 cm of Mazama ash was deposited and ~1.5 cm of St. Helens ash was deposited as airfall. The recurrence interval for such events can be estimated using the Volcanic Explosivity Index (VEI) of Newhall and Self (1982). Mount St. Helens has a VEI of 5 (Newhall and Self 1982) and Mt. Mazama has a VEI of 7 (Ward 2009), implying recurrence time scales of centuries for the former and millennia for the latter. These values would require either about 9 Mt. Mazama-scale events or 27 Mt. St. Helens-scale events per millennium, on average, to explain the change in stacking by itself. Given that much of the ash found in the Brushy Basin floodplain deposits shows evidence of reworking (Bradshaw and Kowallis 2009, 2010), the number of eruptions could have been fewer if much of the ash was transported from surrounding highlands to the flood plain as wash load in rivers.

A second mechanism could have helped greatly reduce stacking density in concert with the mass volume effect described above by impacting avulsion frequency rate. Following the results of Mohrig et al. (2000), we assume that avulsion is driven ultimately by potential energy generated by relief between the channel belt and the floodplain driven by differential aggradation rates (Fig. 12A). The greater the aggradation rate of the channel belt relative to the floodplain (Fig. 12B), the more frequently avulsions take place. Conversely, as floodplain aggradation rate increases due to the addition of ash (Fig. 12C), it takes longer for the channel belt to reach the critical relief needed for avulsion to occur, resulting in less frequent avulsions. This, in turn, decreases stacking density because avulsion is less frequent and, when it does occur, there is thicker flood-plain sediment separating superjacent channel-belt sand bodies.

The specific impact of ash deposition on the floodplain depends upon what the background floodplain aggradation rate is with respect to the channel belt. A simple geometric model, following the approach of Figure 12, where channel aggradation rate (a) is 1 m/ka, the background floodplain aggradation rate (b) is 50 cm/ka, and the critical relief for avulsion is 1 m, shows that if ash deposition rate (c) was also 50 cm/ka, the frequency of avulsion would be reduced by a factor of two, leading to a reduction of stacking density by 65%.

**Comparison with Wasatch Formation**

Our results show that the change in stacking density from the Salt Wash to the Brushy Basin members of the Morrison Formation developed with no documentable change in paleochannels (grain size and flow depth), or in geometry of the channel belts (sinuosity, sandbody thickness, number of stories). There was a pronounced change in composition and abundance of the flood-plain mudstones which went from relatively little evidence of volcanic ash influx to locally abundant, although varying, ash content. Above we suggested that the observed change in stacking density could be explained by the increase in ash influx.

These results contrast with those from a parallel study conducted on the Wasatch Formation (Paleogene) in western Colorado, U.S.A. (Foreman et al. 2012). The Wasatch Formation contains two rapid shifts in stacking density. The floodplain-dominated lower part (the Atwell Gulch Member), rapidly passes into a sandbody-dominated middle part (the Molina Member), which passes rapidly back into a floodplain-dominated upper part (the Shire Member). The lower transition occurs during the time of onset of the Paleocene–Eocene Thermal Maximum (PETM). The PETM lasted c. 200 thousand years during which global temperatures rose, on average 5–8°C, and, at the paleolatitude of the Wasatch Formation, global climate models suggest an increase in annual precipitation by c. 10% relative to early Paleocene conditions (Winguth et al. 2010). As the PETM began, the deposits rapidly changed into higher stacking density (from about 10% to 60% net sandstone). At the same time the depth of the rivers that deposited sand bodies as well as the size of the channel belts increased on average, consistent with an increase in water supply to the system (Foreman et al. 2012). There were attendant changes in types of bedforms that constitute the channel fill, the relative abundance of crevasse-splay deposits, and grain size of the floodplain materials. However, after the end of the PETM, the sedimentary system only slowly returned to its earlier, finer-grained, stratigraphic character, indicating hysteresis in the response of the fluvial system to climate change.

The Morrison Formation is a very different system. It is surprising that such a profound change in stacking density can occur without any attendant change in river or channel-belt geometry. As such, this indicates that very different processes can produce nearly identical stratigraphic characteristics. In neither the Morrison nor the Wasatch formation examples are downstream controls by base level needed to explain the observed changes. Equifinality in the stacking density of both of these examples indicates, at minimum, that simple application of alluvial architectural models, such as originally proposed by Leeder (1978), Allen (1978), Leeder (1978), and Bridge and Leeder (1979) and then modified to become part of the nonmarine sequence-stratigraphy paradigm (Wright and Marriott 1993; Shanley and McCabe 1994), results in nonunique interpretation.

**SUMMARY AND CONCLUSIONS**

Changes in fluvial stacking density is an obvious and ubiquitous characteristic of alluvial stratigraphy. Published models have emphasized the role of varying aggradation rate on changes in channel-belt stacking patterns. The Morrison Formation in eastern Utah and west-central Colorado contains a profound change in stacking density between the Salt Wash and Brushy Basin members. We find that there are no obvious changes in most measured depositional characteristics of paleoriver channels or channel belts between the two units, despite the change in stacking. The major change documented is in the abundance and composition of clay constituting floodplain deposits. Fine-grained floodplain deposits become dominant in the Brushy Basin Member, although apparently rivers were of the same form and size as those found in the Salt Wash Member. This leaves the impression that nothing changes other than that the entire sequence became inflated with floodplain mudstones. Clay composition also changes from illite-dominated to smectite-dominated nearly coincident with the change in stacking pattern. The change in clay composition and abundance indicates an overall increase in volcanic ash delivered to the floodplain either directly as airfall or reworked and transported to the basin as washload in rivers. Geometric modeling demonstrates that the change in stacking density could have directly resulted from the introduction of ash, if the ash made up more than 80% of all floodplain deposits. This volume rate of ash influx could have been supplied by volcanic centers with eruption rates equivalent to Mt. St. Helens or Mt. Mazama.

Our results indicate that the change in stacking pattern observed in the Morrison Formation was coincident with the rapid change in rate of ash supplied to the basin. If this was a threshold event, it was geologically instantaneous with
Appendix

Analysis of Stacking Density in the Rocky Mountains

To quantify the distribution of stacking densities for a variety of fluvial sequences, we collected and analyzed 78 large-scale photo panoramas taken of well-exposed fluvial successions of Jurassic–Paleogene age found in the U.S. Rocky Mountains. These photos were taken from nine fluvial formations exposed in eastern Utah and western Colorado (Fig. A1). In each photo we first mapped the channel-belt and floodplain deposits. We then picked a square window with a side dimension equal to six times the maximum interval thickness of the minority unit (e.g., mudstone if sandstone dominates). This 6 × scale was used because significantly smaller windows converged towards bimodal results (i.e., single pixels are either sandstone or mudstone) and much larger windows tended to reduce all windows towards the average net-sandstone percent for the entire photo, losing information about the heterogeneity of exposures. For each photo panel we then moved the window randomly 500 times, to positions that had at least 80% exposure (i.e., areas that were not sky, covered, or otherwise not interpretable) and calculated the percent net sandstone in each window. The data for each photo is shown as violin plots in Figure 2 (for the Morrison Formation) and Figure A1 (for all other units studied). Violin plots are similar to box plots that include the probability density of the data at different values. Units tend to be either very muddy or very sandy, with 58% of all windows either being 0–20% or 80–100% net sandstone.

Fig. A1.—Net abundance of channel-belt sandstone observed in 26,000 windows taken at random across 52 photo pans of fluvial stratigraphic units in western Colorado and eastern Utah, normalized per formation (i.e., each plot represents 100% of the data for each unit). Units analyzed are: Castlegate Sandstone (Campanian), Colton Formation (Paleocene–Eocene), Farrow Formation (Campanian), North Horn Formation (Maastrichtian–Paleocene), Price River Formation (Campanian–Maastrichtian), Wasatch Formation (Paleocene–Eocene, Molina Member and the Atwell Gulch and Shire members combined), and Williams Fork Formation (Campanian–Maastrichtian). Mean % sandstone for all units is 37%. Ages from Fouch et al. (1982) and Heller et al. (2013).