MULTI-SCALE STRATIGRAPHIC AND STATISTICAL ANALYSIS OF ALLOGENIC AND AUTOGENIC CONTROLS ON FLUVIAL SYSTEMS

by

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ABSTRACT

Fluvial systems are important hydrocarbon reservoirs around the globe, including the high net-sand content fluvial reservoirs of the North Sea and the low net-sand content Mungaroo of offshore Australia. Despite their economic significance, fluvial successions are a challenging reservoir type to predict, characterize, and model. Because large amounts of hydrocarbons are stored in subsurface fluvial reservoirs, understanding the stratigraphic expression of external (allogenic) and internal (autogenic) forcing mechanisms at multiple scales is key to predicting reservoir connectivity from the large basin to the small bar scale.

At the basin scale this dissertation quantitatively compares and contrasts the influence of lateral boundary conditions on fluvial channel belt stacking patterns. Specifically, how the valley confined Dakota Sandstone is inherently different than the unconfined lower Wasatch Formation in regards to clustering, compensational stacking, and connectivity (Chapter 2). Results from this chapter document that the confined Dakota Sandstone has stronger clustering, lower compensation stacking and higher connectivity than the unconfined lower Wasatch Formation. However, both systems show similar longitudinal trends in these characteristics.

At the channel belt scale (Chapter 3), this dissertation puts forth process-based theory coupled with satellite, seismic, outcrop, and numerical experiments to document how the autogenic morphodynamics of the sediment routing system control the planform shape of channel belts. Specifically, that the erosion coefficients of the subjacent and lateral material determine the final channel-belt morphology given long enough residence time on the floodplain.

At the smallest spatial scale, the bar scale (Chapter 4), this dissertation uses facies proportions and sedimentary structures coupled with a paleomorphodynamics workflow to document persistence or transience of mean flow velocity. This in turn was used to infer
perennial and ephemeral flow conditions. Furthermore, this chapter documents that allogenic signals can either be preserved or shredded by the stratigraphic filter depending on the accretion style of the channel belts. Intra channel-belt signal preservation comes at the expense of basin-scale preservation, where either channel stacking patterns or barforms record the allogenic signal but not both.
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CHAPTER 1

INTRODUCTION AND DISSERTATION FORMAT

This is the introduction to the dissertation. Herein, I discuss general background on fluvial systems, and the format of this dissertation. This chapter is divided into three parts. First, background information on fluvial system, followed by the organization and topics of each chapter, and finally references.

1.1 Introduction to Fluvial Systems

The stratigraphic expression of fluvial systems is important from both scientific and resource extraction perspectives. Scientifically, fluvial systems provide insights into past environments on earth. A large portion of human population lives on or near active river channels which makes understanding how fluvial systems evolve important. From a resource extraction perspective, fluvial strata host large amounts of hydrocarbons, which is an important economic factor for many countries. Despite their economic significance, fluvial successions are a challenging reservoir type to characterize and model (Shepherd, 2009; Pranter and Sommer, 2011). Advances in modern 3D seismic have made imaging fluvial successions much easier, but attenuation of seismic waves at reservoir depth results in resolution of only 10’s of meters. This coarse resolution does not provide insight into the internal architecture, such as facies distributions and bed styles, or the interconnectedness of sandstones between adjacent channel belts, or the reservoir properties of fluvial channel belts. Well logs and cores provide high-resolution documentation of the internal characteristics of channel belts such as architecture and facies. However, they only provide a 1D vertical profile through the channel belt (Shepherd, 2009). As such, predicting the connectivity within channel belts at the bed scale in higher
dimensions proves problematic in the subsurface. However, well-exposed outcrops provide high-resolution 2 and 3-dimensional exposures that can be used in reservoir modeling and prediction.

1.2 Dissertation Format

This dissertation is composed of three Chapters that have been submitted to peer-reviewed journals, and are described in detail below. Each chapter is formatted according to the journal to which it is submitted. Therefore, each chapter will contain a separate abstract, introduction, geologic setting, data and methods, discussion, conclusion, and references.

In this dissertation, I use a multi-scale statistical process-based stratigraphic method to quantitatively document fluvial systems in 3 distinctive but broadly integrated studies.

• **Chapter 2-** Chapter 2 deals with how fluvial channel belts move over million year time scales at the basin scale (100’s of km) and allogetic controls. Specifically, how lateral boundary conditions influence channel-belt stacking patterns in a down-current transect (Figure 1.1a). Furthermore, Chapter 2 emphasizes how channel belts are connected to one another in 3 dimensions. Understanding how lateral boundary conditions influence channel-belt stacking patterns and connectivity for unconfined and confined is important for optimizing well placements for oil and gas production.

• **Chapter 3-** Chapter 3 of this dissertation concentrates on shorter time scales (100-500 years) and spatial scales (100’s of meters) and autogenic controls on the shape of ancient and modern river channels. This chapter deals with how the shape of channel belts are controlled by the hydrodynamics within the active channel as it migrates across the landscape (Figure 1.1b). This part of my research again is critical for oil and gas exploration as 3D seismic can image channel belts, but not the internal characteristics
(e.g. facies, bar accretion type) within the channel belt. Furthermore, this chapter is important for modern day land use around modern rivers.

- **Chapter 4-** Chapter 4 focuses on the shortest time scales (seconds-100 years) and spatial scales (microns-meters) associated with grain-to-grain interactions. This Chapter quantitatively documents the spatial location of facies within 5 selected channel belts. Additionally, this Chapter uses grain size and a paleomorphodynamic workflow to document how the hydrodynamics within the channel belt changed over these short time scales (Figure 1.1c). This is important for two reasons, (1) it provides a quantitative method to model channel belts for reservoir characterization, and (2) it documents how flow processes changed within channel belts without being able to directly measure it.

- **Chapter 5-** This final chapter discusses how Chapters 2-4 have increased our scientific understanding of fluvial systems at different temporal and spatial scales.

### 1.3 References


Figure 1.1. Comparison of the different spatial scales of fluvial systems documented in this dissertation. (A) A schematic cross section through a basin showing channel belt stacking patterns. The channel belt in (B) is outlined by the dashed rectangle. (B) Schematic 3D diagram of a meandering fluvial channel belt showing channel belt morphology and channel sinuosity. The cross-section in part (C) is outlined by the dashed rectangle. (C) Schematic diagram of the intra channel-belt facies within the channel belt in part (B) documenting bedset scale facies changes.
CHAPTER 2

THE INFLUENCE OF LATERAL BOUNDARY CONDITIONS ON FLUVIAL CHANNEL-BELT CLUSTERING, COMPENSATION AND CONNECTIVITY: LOWER WASATCH FORMATION AND DAKOTA SANDSTONE, UTAH

A paper to be submitted to Sedimentology

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2.1 Abstract

Fluvial channel-belt clustering has recently been documented using quantitative metrics for systems dominated by autogenic controls. Furthermore, it has been long recognized that allogenic forcing (tectonic and eustatic controls) can lead to confinement of fluvial systems, resulting in clustering of channel belts. However, to date no study has quantitatively documented the differences in channel-belt clustering, compensational stacking of channel belts, and intra channel-belt connectivity in unconfined and confined systems. Herein, we quantitatively compare world-class outcrops of an unconfined fluvial system (Paleocene lower Wasatch Formation), with outcrops of a confined fluvial system (Cretaceous Dakota Sandstone). Furthermore, we develop two new methods to quantitatively document channel-belt clustering and intra channel-belt connectivity. Using these new methods, and other previously developed methods, we document an increase in channel-belt clustering and intra channel-belt connectivity down dip in both systems. Additionally, we find that channel belts within the unconfined system stack more compensationally than those in the confined system. These new methods and empirical relationships can be applied to fluvial reservoir prediction and modeling. The
workflows are important for predicting intra channel-belt connectivity, and accurately modeling
unconfined and confined fluvial systems in the oil and gas industry.

2.2 Introduction

Compensational stacking and clustering are important measures of stratigraphic
architecture in fluvial successions (Straub et al., 2009; Hajek et al., 2010). Furthermore, connectivity between fluvial channel belts is an important measure for understanding how fluids move through subsurface reservoirs. However, to date no study has used quantitative methods to relate compensational stacking and clustering to connectivity. Herein, we use previously developed metrics, modified methods, and newly developed methods to quantitatively relate compensational stacking and clustering to connectivity in fluvial stratigraphy.

Channel-belt stacking patterns describe how channelized deposits are spatially located relative to one another (Allen, 1978; Clark and Pickering, 1996; Straub et al., 2009). Two methods used to describe stacking patterns in fluvial systems are: (1) clustering and (2) compensational stacking. Clustering refers to the groupings of channel-belts relative to each other, whereas compensational stacking refers to the tendency of a sediment transport system to fill topographic lows (Straub et al., 2009; Hajek et al., 2010). An additional measure, connectivity, is needed to document interconnectedness of sandstones in adjacent channel belts. This measure directly relates to reservoir connectivity, which refers to the amount of a hydrocarbon reservoir that is connected in a given volume (Larue and Hovadik, 2006). Whereas clustering, compensational stacking and connectivity studies have been independently conducted in autogenic dominated systems (e.g. Straub et al., 2009; Hajek et al., 2010; Funk et al., 2012) the relationships between these parameters, nor the cross-sectional and plan-view shape of clusters, has not yet been documented for both autogenic and allogenic dominated systems.
This study broadly categorizes fluvial systems into two categories based on their degree of lateral confinement relative to channel-belt size: confined and unconfined. Confined systems in the modern are recognized by two primary characteristics: (1) their plan form geometries are overall rectilinear elongate features at the basin scale, and (2) the lateral extents of the system are controlled by laterally adjacent topographic highs, such as valley walls, generated during sea-level fall, or in uplifting terrains, or tectonically generated topography such as a fault scarps (Figure 2.1b). Whereas the forcing mechanisms have been hotly debated (Holbrook, 2010; Strong and Paola, 2010), we simply group all confined systems on the basis of their geometric patterns. Previous workers have defined valley confined systems as incised valleys, rift confined systems, and longitudinal systems in compressional terranes (e.g. Van Wagoner et al., 1990; Dalrymple et al., 1994; Posamentier and Allen, 2001; Strong and Paola, 2008; Bhattacharya, 2011; Holbrook and Bhattacharya, 2012; Blum et al., 2013; Li and Bhattacharya, 2013).

Unconfined systems are recognized in the modern by two primary characteristics: (1) their large-scale plan form geometries are radially dispersive in shape, and (2) the only topographic relief within the system is generated by channel belts as they avulse across the floodplain (Figure 2.1a). Unconfined systems have been called fluvial megafans, distributive fluvial systems, terminal fans, and losimean fans (e.g. Geddes, 1960; Kumar, 1993; Kelly and Olsen, 1993; Singh et al., 1993; Stanistreet and McCarthy, 1993; Sinha and Friend, 1994; Gupta, 1997; Nichols and Hirst, 1998; DeCelles and Cavazza, 1999; Horton and DeCelles, 2001; Leier et al., 2005; Hartley et al., 2010; Weissmann et al., 2010; Buehler et al., 2011; Davidson et al., 2013). This study focuses on exceptionally well exposed outcrops of the unconfined lower Wasatch Formation, and a set of world class exposures of the valley confined Dakota Sandstone (Figure 2.2).
Previous studies have documented that lateral boundary conditions present during deposition (allogenic controlled) have strong impacts on channel-belt density and net-sand content (Bridge and Leeder, 1979). The goal of this study is to quantify the following: (1) the relationship between clustering and compensation, (2) the relationship between connectivity and clustering, and (3) the plan-view shape of clusters in confined (allogenic) and unconfined (autogenic) fluvial systems. This research is important because fluvial systems are significant hydrocarbon reservoirs around the globe, including the high net-sand content fluvial reservoirs of the North Sea (Labourdette and Jones, 2007) and the low net-sand content Mungaroo of offshore Australia, and the Kern River field of California (Shepherd, 2009; Stoner, 2010). Despite their economic significance, fluvial successions are a challenging reservoir type to predict, characterize, and model (Shepherd, 2009; Pranter and Sommer, 2011). Seismic and well data are used to constrain the geometry of fluvial reservoirs, however, fluvial architecture and connectivity is often below seismic resolution and well logs provide only a 1D profile (Shepherd, 2009). Because large amounts of hydrocarbons are stored in subsurface fluvial reservoirs, understanding how connectivity and channel-belt stacking patterns relate to lateral confinement is critical to creating accurate subsurface models that can be used to maximize economic viability of these fields.

2.3 Geologic Setting

The unconfined lower Wasatch Formation and confined Dakota Sandstone are two end members in the spectrum of confined-to-unconfined (respectively) boundary conditions. The Wasatch Formation of the Uinta Basin in eastern Utah contains excellent exposures of a low net-sand content unconfined fluvial system (Ford and Pyles, 2014). The Uinta Basin is a longitudinally asymmetric, foreland basin located in northeastern and central Utah in the core of
the Laramide structural province (Figure 2.2a). The 24,000 km² basin is bounded on the north, east, and south by the Uinta uplift, Douglas Creek arch, and Uncompahgre uplift, respectively (Figure 2.2b). The San Rafael uplift and Sevier fold and thrust belt delineate the western margin of the basin (Dickinson et al., 1988; Montgomery and Morgan, 1998). From the Paleocene through the Eocene, flexurally induced subsidence within the Uinta basin provided accommodation for deposition of lacustrine strata of the Green River Formation and fluvial strata of the Wasatch Formation. Lacustrine sediments are surrounded by deltaic and fluvial strata of the Green River and Wasatch Formations; signifying internal drainage (Picard, 1955; Keighley et al., 2002).

Paleocurrent directions in the southern outcrops of the Wasatch Formation document that the fluvial systems flowed north and northeast towards the center of the basin (Dickinson et al., 1986; Dickinson et al., 2012; Ford and Pyles, 2014). The Wasatch Formation is divided into three units based on net-sand content, each bounded by compound paleosols: the lower, middle, and upper Wasatch (Ford and Pyles, 2014; Sendziak, 2012). The low net-sand content fluvial deposits of the unconfined lower Wasatch Formation are the focus of this study. The unconfined lower Wasatch Formation outcrops are located along the southern margin of the Uinta basin west of the modern Green River, and have excellent exposures of channel belts, crevasse splays, and floodplain fines. Three strike-oriented outcrop exposures located in successively down paleodip positions were used to address the goals of this study (Figure 2.3). The lower Wasatch outcrops range from 5 to 2.5 km wide, are 300 m tall, and span 3 km longitudinally down paleodip.

The Cenomanian Dakota Sandstone of the now dissected Sevier basin in central Utah contains an exceptionally well exposed valley-confined fluvial system deposited near the crest of the Sevier forebulge (Kirschbaum and Schenk, 2010). The longitudinally asymmetric, laterally
continuous 2,500,000 km² Sevier basin was bounded to the west by the Sevier fold and thrust belt and to the east by the intercontinental arch (DeCelles and Coogan, 2006). The forebulge of the basin during the Cenomanian was located in the area now occupied by the San Rafael Swell. This is documented by visible thinning of Cenomanian strata eastward over the Swell, and thickening of the strata to the west into the Sevier basin foredeep (DeCelles, 2004). It unconformably overlies the Cedar Mountain Formation, and is conformably overlain by the Tununk member of the Mancos Shale (Kirschbaum and Schenk, 2010).

The Dakota Sandstone in the study area has been interpreted by Kirschbaum and Schenk (2010) to be an exceptionally well preserved fluvial system confined within a valley. The preservation of dominantly fluvial channel belts is unique as most incised valley systems are typically filled with estuarine deposits (Kirschbaum and Schenk, 2010). Channel belts are composed primarily of downstream accreting bars in the lower Dakota, while in the upper portion channel belts contain a balance of downstream and laterally accreting bars. Paleocurrent measurements document that the fluvial systems flowed to the northeast (Kirschbaum and Schenk, 2010).

The confined Dakota Sandstone outcrops are located along the western limb of the San Rafael Swell, near Mesa Butte, and have excellent exposures of channel belts, crevasse splays, and floodplain fines. Three strike-oriented outcrop exposures located in successively down paleodip positions were used to address the goals of this study (Figure 2.4). The Dakota outcrops range from 1 to 2 km in width, are 20 m tall, and span 2.5 km longitudinally down paleodip. The apparent channel-belt clustering, excellent exposure, and large size of both outcrops provides a natural laboratory to quantify clustering, connectivity, and channel belt stacking patterns in both unconfined and confined fluvial strata. For both the confined (Dakota Sandstone) and unconfined
(lower Wasatch Formation) systems, the cross sections are spaced 1 to 2 km down dip from one another.

Whereas the depositional setting of the lower Wasatch Formation and Dakota Sandston vary (unconfined and confined) there are notable similarities in terms tectonic setting, paleoclimate, duration of deposition, primary channel-belt facies, vertical changes in bar accretion types, and channel-belt aspect ratios (Table 2.1). The similarities between the two systems at the time of deposition further provides constraints that the major difference between the systems is the degree of lateral confinement.

2.4 Dataset and Methods

The following data was used to address the questions of this study: 1) half-meter to decimeter resolution measured sections documenting lithofacies, 2) high-resolution photo panels, 3) paleocurrent measurements, and 4) laser rangefinder measurements. Geologic maps, GPS points, and photo panels were used to document features walked out in the field, while laser range finding was used to constrain spatial positions and dimensions of the data. These initial data were collectively used to generate additional information about the outcrop: 1) spatial locations of channel belts and their centroids, and 2) channel-belt orientations, and 3) channel belt dimensions.

The influence of lateral boundary conditions on channel-belt stacking patterns was investigated in context of clustering, compensational stacking, and connectivity of channel belts. Clustering was analyzed using the following: Ripley’s K function, point density maps, and a new method develop herein based on the Manhattan Distance ($R3$). Compensational stacking was evaluated using the modified compensation index ($K_{cm}$). Connectivity was evaluated using a new methodology developed herein referred to as the Gamma Index. Each method is discussed below.
and applied to the both of the cross sections in the confined (Dakota Sandstone) and unconfined (lower Wasatch Formation) study areas.

Clustering of channel belts, describes the spacing of channel-belts relative to each other (Hajek et al., 2010). Hajek et al. (2010) used Ripley’s K function to quantify clustering of the centroids of fluvial channel belts in the Ferris Formation of southern Wyoming. Ripley’s K function is defined as:

$$\hat{K}(h) = \hat{\lambda}^{-1} \sum_{i=1}^{N} \sum_{j=1}^{N} w(s_i, s_j) \frac{I(\|s_i - s_j\| \leq h)}{N}, \quad h > 0$$

where \(w(s_i, s_j)\) is a weighting factor or edge correction that is equal to one when the circumference of a circle centered at the point \(s_i\) and passes through \(s_j\) is completely within the bounds of the study area \(A\), and is a proportion when part of the circle falls outside the study area, and \(I(x)\) is the indicator function. Ripley also noted that for the edge correction to be statistically significant \(h\) should be less than one half the shortest length of the study area (Cressie, 1991; Dixon, 2002).

Ripley’s K function is a powerful method to quantify clustering but at least three potential shortcomings exist. First the Ripley K function is a stationary point process designed for isotropic media such as stars, or trees when viewed on a map, and fails to address clustering of anisotropic geometric shapes that occupy two dimensional space (\(R^2\) space), such as the cross sections of channel belts. As such the approach by Hajek et al. (2010) is problematic. Second, the length of the shortest side of the study area limits the statistical range of the K function to distances much shorter than the study area as a whole. For example if a dataset is 300 m thick and 5000 m wide, the K function will only allow for clusters up to 75 m in diameter. Third, the Ripley K function does not identify direct connections between anisotropic shapes in \(R^2\). Despite
these potential shortcomings, we use this metric to maintain continuity with earlier articles that used this approach (e.g. Hajek et al., 2010).

To supplement Ripley’s K function, we use an approach developed by Diggle (1985), which is an isotropic Gaussian kernel smoothing function for point data. This method is used to contour point densities for spatial data. Herein, the point density approach of Diggle (1985) is used to document the size, shape, and locations of channel-belt clusters in both the confined and unconfined systems, both in focus of cross-sections and in map view.

To address the shortcomings of Ripley’s K function discussed above for clustering, we developed a new method for documenting channel-belt clustering, which is modified from Sheets (2004). Sheets (2004) developed a probability distribution function of the Euclidian Distance between all pairwise combinations of the channel bases to quantify channel stacking patterns. This method is robust for quantifying the average distance between channel bases and documents multiple levels of clustering in synthetic datasets. However the method developed by Sheets (2004) does not discriminate lateral from vertical stacking, nor does it account for the anisotropy of channel belts in cross section (ie – widths and thicknesses). To account for these shortcomings, we revise Sheets (2004) method. The new method is based on the component vectors of the Manhattan Distance ($d$), which is the sum of the absolute differences of two centroids, and is defined as

$$d ( p, q ) = \| p - q \| = \sum_{i=1}^{n} | p_i - q_i |$$

(2)

where $p$ and $q$ are two component vectors in two dimensional space ($\mathbb{R}^2$) with Cartesian coordinates ($p_1, p_2$) and ($q_1, q_2$), respectively. The vector lengths are then normalized by average channel dimension (width and thickness).
\[ X_{\text{norm}} = \frac{\Delta x}{\bar{x}}, \quad Y_{\text{norm}} = \frac{\Delta y}{\bar{y}} \]  

(3)

where \( \Delta x \) is the horizontal offset between channel belt centroids, \( \Delta y \) is the vertical offset between channel belt centroids, \( \bar{x} \) is the average channel belt width, and \( \bar{y} \) is the average channel belt thickness. \( X_{\text{norm}} \) and \( Y_{\text{norm}} \) are dimensionless, and allows systems with large channels to be compared to those with smaller channels. 2D probability density functions of the horizontal and vertical Manhattan distances between channel centroids for all pairwise combinations (Figure 2.5a,b) documents the distribution of \( d \). This modified method is herein termed the R3 method for clustering. Peaks in the histogram document clusters, or repeated lateral and vertical offsets of channel belts (Figure 2.5). A R3 histogram with peaks close to 0 indicates closely spaced, amalgamated channel belts, whereas peaks greater than 1 indicate spacing greater than 1 channel belt width and thickness (Figure 2.5c). Complete spatial randomness creates histogram that decreases at the same rate in the x and y dimensions and has pronounced anomalies. However regularly spaced points result in a histogram with regularly spaced peaks at set distances.

“Compensational stacking is the tendency of a sediment transport system to preferentially fill topographic lows” (Straub et al., 2009, p. 673). Over short time intervals, the deposition of sediment is controlled by the morphodynamics of the sediment transport system and the local topography (Straub et al., 2009). Over larger time intervals however, the sediment transport system has a higher likelihood of depositing sediment over a wider area within the basin (Straub et al, 2009; Straub and Pyles 2012).

Herein, we used the modified compensation developed by Straub and Pyles (2012) to quantify compensational stacking in outcrops:

\[ CV = \left( \int_L \left[ \frac{\Delta y(x, \Delta x, \Delta y)}{\Delta y_{\lambda, \mu}} \right]^2 dL \right)^{1/2} \]  

(6)
where $A$ and $B$ are stratigraphic surfaces, $\Delta \eta(x)_{A,B}$ is the average local deposit thickness between surfaces $A$ and $B$ measured over the cross section $L$. The modified compensation index $(\kappa_{CV})$ is the power law decay trend as average stratigraphic thickness increases:

$$CV = a \Delta \eta_{A,B}^{-\kappa_{CV}}$$  \hspace{0.5cm} (7)$$

where $CV$ is equal to the product of a leading coefficient and local deposit thickness between two stratigraphic surfaces, and $\kappa_{CV}$ is the covariate of compensation and, in natural systems, ranges from 0 to 1. $\kappa_{CV}$ values ranging from 0.0 to 0.5 indicate clustering of deposits, values of 0.5 indicate random stacking patterns, and values of 1.0 correlate to pure compensational stacking patterns (Straub et al, 2009; Straub and Pyles 2012).

Despite the advances in quantitative description of channel-belt stacking patterns, little quantitative data exists that relates large-scale stacking patterns to channel-belt connectivity. The term connectivity is used herein as the fraction of channel belts that are either directly connected to one another or indirectly through another channel belt. Allen (1978) numerically described how fluvial channel belts connect to one another by documenting the average fractional contact between all sides of the channel belts. Allen (1978) used residual area density as the total area of sand bodies preserved in cross section divided by the entire cross sectional area. Leeder (1977) used a similar approach to describe channel belts and floodplain deposits, and his model took into account sinuosity of channel belts. Leeder (1977) devised a probability function to describe how channel belts can contact one another based on differential compaction. Leeder (1977) also devised an interconnectedness ratio that described the ratio of touching channel-belts to non-touching channel-belts, but neither method takes into account higher order connectivity of channel belts to one another through intermediate channel belts. The method of connectivity for
deepwater channels proposed by Funk et al. (2012) is similar to the connectivity formulas of Leeder (1977) and Allen (1978) but uses the fractional length of sand-on-sand contacts between adjacent channels. Although the method of Funk et al. (2012) is useful for facies-scale connectivity, it is much more difficult to apply globally to large-scale outcrops. To address this issue, herein a new method is developed to describe connectivity. The new method describes connectivity ($\gamma$) as a function of the shortest path distance between centroids

$$z = \frac{L_{CB}}{L_{Total}}$$

(8)

where $L_{CB}$ is the length of the Bellman-Ford shortest path between two channel belt centroids that is within channel-belt elements, and $L_{Total}$ is the total Bellman-Ford shortest path distance between two channel belt centroids (Bellman, 1956). Using this definition of $z$, we then define connectivity ($\gamma$) as

$$\gamma = \frac{n_{z=1} + n_c}{n^2 - n}$$

(9)

where $n_{z=1}$ is the number of pairwise combinations of channel belts where the shortest path between two centroids is contained entirely within channel belts, $n_c$ is the number of connected channel belts, and $n$ is the number of channel belts being evaluated in the algorithm. $Z$ ranges from 0 to $\infty$ where 0 indicates that no channels are connected and values greater than 1 document that all channels are connected. High values of $Z$ correspond to a R3 histogram with peaks between 0 and 1, whereas low values of $Z$ correspond to a R3 histogram with peaks larger than 1. Using the $\gamma$ index to describe connectivity is beneficial because it not only documents connectivity of adjacent connected channel belts, but also connectivity of channel belts that are connected through intermediate adjacent channel belts. Furthermore, this method documents element-scale connectivity and is based on geometries and spatial locations of channel belts.
which is useful for understanding large-scale connectivity between channel belts. This is similar to the method used by Hovadik and Larue (2007) but does not use channel belt volumes.

2.5 Results

Herein, we present the results for channel-belt clustering, compensational stacking, and connectivity for both the unconfined lower Wasatch Formation and confined Dakota Sandstone from updip-to-downdip. We then compare and contrast the two end-members in the spectrum of lateral confinement.

2.5.1 Clustering

Three methods were used to document channel-belt clustering in the confined and unconfined systems (Ripley’s K, R3, point density maps). The results are summarized below. Within the unconfined lower Wasatch, Ripley’s K function documents that the minimum clustering distance decreases slightly from 45 to 40 m before increasing to 65 m longitudinally down paleo-dip (Figure 2.6A). In contrast, within the confined Dakota, Ripley’s K function documents a decrease in minimum clustering distance longitudinally from 11 m to 3 m down paleo-dip (Figure 2.6).

For the unconfined Wasatch system the R3 method documents a change from predominantly vertical stacking of channel belts in the most updip outcrop to both lateral and vertical stacking of channel belts in the most down dip outcrop (Figures 2.7 and 2.8). Furthermore, the R3 method documents that most channel belts are located within 5 average channel belt widths and thicknesses at all positions. However, the R3 method documents a large distribution of distances between channel belts, both vertically and laterally in the unconfined lower Wasatch Formation that changes from from laterally stacked channel belts to more vertically stacked channel belts in the down-current direction, with most channel belts being located within 5 average channel-belt
widths laterally, and 2.5 channel-belt thicknesses vertically (Figures 2.7 and 2.8). However, the distribution of distances between channel belts is much smaller in the confined Dakota Sandstone than in the unconfined lower Wasatch Formation.

Point density maps (Figure 2.9) were used to constrain cluster widths, thicknesses, and minimum lengths. Cluster widths in the unconfined lower Wasatch Formation decrease from a mean of 321 m in the up-dip outcrop to a mean of 105 m for the medial outcrop, and 130 m for the down dip outcrop (Figure 2.10a). Similarly, cluster thickness decreases from a mean of 124 m in the up dip outcrop to 84 m for the medial outcrop, and 81 m for the down dip outcrop (Figure 2.10a). Cluster aspect ratios decrease as well, from a mean of 2.5 for the up dip outcrop to a mean of 1.3 for both the medial and down dip outcrops (Figure 2.10b). Cluster lengths in the unconfined system range from 360 m to 3,080 m with a mean of 1,461 m in length (Figure 2.10b). We acknowledge that these are minimum lengths, as the outcrop dataset is limited in the up and down-current direction. Therefore, the cluster lengths can be significantly longer than measured in the outcrop datasets.

Cluster widths in the confined Dakota Sandstone document an inverse trend to clusters in the unconfined lower Wasatch Formation. In the confined Dakota, cluster widths increase down paleo-dip from a mean of 28 m in the up-dip outcrop, to a mean of 48 m in the medial outcrop, and finally to a mean of 187 m in the down-dip outcrop (Figure 2.10a). Cluster thicknesses in the confined Dakota follow a similar trend, with means of 20 m, 17 m, and 19 m in up-dip, medial, and down-dip respectively (Figure 2.10a). These patterns result in a down-dip increase in cluster aspect ratios with means of 1.3, 2.6, and 9.6 for up dip, medial, and down dip respectively (Figure 2.10b). Cluster lengths in the confined Dakota system are much smaller than those in the
unconfined lower Wasatch system with a minimum of 195 m to 1500 m, and a mean of 606 m in length (Figure 2.10b).

Building on cluster width and thicknesses, we compare cluster area to number of channel belts per cluster (Figure 2.11). In both systems we document that as cluster area increases, the number of channel belts per cluster increases following a linear trend (Figure 2.11). Meaning that as cluster cross-sectional area increases there are more channel belts within the cluster. Therefore, larger clusters have more channel belts within them than their smaller counterparts.

All these measures of clustering document that there is indeed significant clustering in both the unconfined and confined systems from up-to-down dip, but that the confined Dakota Sandstone has shorter distances between channel belts than the unconfined lower Wasatch Formation. Additionally, the R3 method better documents spatial trends than the Ripley’s K function, probably because the method takes channel-belt widths and thicknesses into account along with the lateral and vertical offsets rather than just absolute distance.

### 2.5.2 Compensation

To complement the point-based clustering metrics that treat channel belts as points, we use the surface based compensation index ($\kappa_{CV}$) to document differences in compensational stacking between the unconfined lower Wasatch and confined Dakota. Specifically, in both the unconfined Wasatch and confined Dakota systems $\kappa_{CV}$ decreases down dip. However, the confined Dakota system has significantly lower $\kappa_{CV}$ values at all transects than the unconfined lower Wasatch System. Meaning the unconfined lower Wasatch channel belts stack more compensationally than their confined counterparts. In the unconfined lower Wasatch system $\kappa_{CV}$ decreases from 0.93 to 0.81 to 0.74 in the down-dip direction (Figure 2.6), documenting a
decrease in compensational stacking. Similarly, the confined Dakota system documents a decrease from 0.85 to 0.70 to 0.68 in the down dip direction (Figure 2.6).

2.5.3 Connectivity

The global $\gamma$ connectivity index documents that both the unconfined and confined systems increase in connectivity down dip (Figure 2.6). The unconfined lower Wasatch Formation has 0.54% of channel belts connected in the up dip outcrop, 1.05% of channel belts connected in the medial outcrop, and 1.72% of channel belts connected in the down dip outcrop (Figure 2.6). In comparison the confined Dakota Sandstone has an order of magnitude greater connectivity, where 3.70% of channel belts connected in the up dip outcrop, 5.88% of channel belts connected in the medial outcrop, and 6.25% of channel belts connected in the down dip outcrop (Figure 2.6).

2.5.3 Synthesis

Channel belts in the unconfined lower Wasatch Formation are less clustered than their counterparts in the confined Dakota Sandstone. Furthermore, compensational stacking of channel belts is higher in the unconfined lower Wasatch Formation than in the confined Dakota Sandstone. Finally, the unconfined lower Wasatch Formation has lower connectivity values than the confined Dakota Sandstone by an order of magnitude.

Longitudinally, the unconfined lower Wasatch Formation has a documented decrease in compensational stacking and cluster width and thickness in a down-current direction. Furthermore, the unconfined lower Wasatch has a documented increase in clustering and connectivity in the down-current direction (Figure 2.12). The confined Dakota Sandstone has a decrease in compensational stacking, but an increase in average cluster width in the down-current direction while cluster thickness remains constant. Additionally, the confined Dakota
Sandstone has a documented increase in clustering and connectivity in the down-current direction (Figure 2.12).

2.6 Discussion

Quantitative methods for evaluating clustering, compensation, and connectivity document some key differences between stratigraphic stacking patterns in the confined Dakota Sandstone and unconfined lower Wasatch Formation. First, $\kappa_{CV}$ and minimum clustering distance, using Ripley’s $K$ function, are weakly related (Figure 2.13). The unconfined system follows a logarithmic decay, while the confined system follows a logarithmic increase. Second, connectivity is weakly related to the minimum clustering distance (Figure 2.13). As the minimum clustering distance increases connectivity decreases for the confined system, and increases for the unconfined system following a linear trend (Figure 2.13). Third, connectivity is directly related to compensational stacking (Figure 2.13). As the compensational stacking ($\kappa_{CV}$) increases with increasing compensational stacking of channel belts, connectivity ($\lambda$) within both the confined and unconfined systems decrease following a power law (Figure 2.13).

The minimum clustering distance calculated by Ripley’s K function (Figure 2.6) has no clear trends from up to down dip in the unconfined and confined systems. We attribute this to the anisotropic shape of channel belts, specifically that their geometric shapes limit how close channel belt centroids can be located to one another. Similarly, the weak trend between Ripley’s K function and the coefficient of variation in Figure 2.13 corroborates the limited use of Ripley’s K function for documenting clustering of channel belts. If minimum clustering distance was directly related to compensational stacking of channel belts, a much stronger correlation would exist. That is, as compensational stacking increases, the minimum clustering distance should
reach a steady state and not increase significantly, as the channel belts should not be able to stack any further apart laterally.

Cluster dimensions can be used as a proxy for strength of lateral confinement. Large clusters have fewer connected channel belts, while small clusters contain more amalgamated well connected channel belts (Figure 2.11). The decrease in cluster dimensions in the down-dip direction for both systems results in better connectivity between channel belts. Critically, we propose that the shape of channel-belt clusters is directly controlled by the lateral boundary conditions (confined or unconfined) present at the time of deposition. Clusters in confined systems have significantly higher aspect ratios than those in their unconfined counterparts. As confinement decreases, clusters reach an aspect ratio of close to 1.

We document for both the confined Dakota Sandstone and unconfined lower Wasatch Formation that as clusters become larger in cross-sectional area, the number of channel belts in a cluster increases linearly. The clusters in the confined Dakota are smaller in cross-sectional area, and have fewer channel belts than the larger unconfined Wasatch clusters. We hypothesize that the lateral confinement limits the size of the clusters, forcing clusters to become only as large as the valley in which the channel belts are confined. The unconfined Wasatch in contrast has a larger depositional area over which channel belts can move, allowing clusters to become much larger while channel belt dimensions remain constant. More work is needed to investigate if this linear trend holds for systems with significantly larger channel belts and clusters.

For the unconfined lower Wasatch, the R3 method in Figure 2.8 documents a down-dip decrease in both lateral and vertical distances between channel belts. This means that the channel belts are more closely spaced in the down dip outcrop, both laterally and vertically, than they are in the up dip outcrop. We note that the majority of channel belts stack within 5 average channel-
belt widths and thicknesses for all three outcrops. However, there is less spread in the
distribution in the down dip outcrop than in the up dip outcrop, which is consistent with
documented $\kappa_{CV}$ values and increasing connectivity ($\gamma$ index) values.

For the confined Dakota, the R3 method in Figure 2.8 documents remarkably consistent
stacking of channel belts from up dip to down dip outcrops. The spread of the distribution does
increase in the down dip outcrop, but again the majority of channel belts stack within 3 average
channel belt widths, and 1 average channel belt thickness in all three outcrops. We interpret this
pattern to document less longitudinal variability of channel-belt stacking patterns in the confined
system than in the unconfined system.

Data in Figure 2.6 documents the decrease in compensation from up to down dip for both
the lower Wasatch Formation (unconfined system), and the Dakota Sandstone (confined system).
This is similar to the longitudinal trends documented by Straub and Wang in an experimental
delta (Straub and Wang, 2013). However, in our case we hypothesize that the lateral boundary
conditions (valley walls) in the confined Dakota Sandstone have lowered the $\kappa_{CV}$, meaning that
in addition to the water-to-sediment flux ratio documented by Straub and Wang (2013),
confinement also has a first-order control on channel stacking patterns. Thus, some allogenic
forcing mechanisms such as confinement have an imprint in the stratigraphy. The controls on the
decrease in $\kappa_{CV}$ in both the unconfined and confined systems are not well understood, however
we interpret this trend to be related to floodplain aggradation rates relative to channel migration
rates (Straub, Pers. Comm.), meaning that down-dip enhanced floodplain aggradation rates cause
the channel migration rates to decrease, resulting in fixed locations of the channels. In other
words, the channels form a contributory pattern resulting in vertical, rather than lateral stacking.
In both the confined and unconfined systems the accommodation within the basins at the time of
deposition is interpreted to increase down dip, as the maximum subsidence during deposition
was located down dip from the location of the outcrop exposures, resulting in a longitudinal
increase in aggradation of the fluvial systems. While this explanation is plausible, we
recommend future studies to further address this pattern.

Connectivity increases in the down-dip direction for both the unconfined lower Wasatch
and the confined Dakota (Figure 2.6), however, the increase is much larger in the unconfined
system. The increase in channel-belt connectivity is directly related to the decrease in the
compensational stacking ($\kappa_{CV}$) and cluster dimensions (Figure 2.13). As the compensational
stacking ($\kappa_{CV}$) decreases down dip, channel belts become more closely spaced, and become
better connected. The empirical relationship between connectivity and compensational stacking
follows a power law decay:

$$\gamma (\kappa_{CV}) = a \kappa_{CV}^{-\lambda}$$ (10)

where $a$ is a leading coefficient, $\kappa_{CV}$ is the coefficient of variation, and $\lambda$ is the degree of lateral
confinement and ranges from 0 for completely confined systems, to $\infty$ for completely unconfined
systems.

Confined systems have a much slower decrease of connectivity with increasing
compensational stacking than unconfined systems. This relationship sets the minimum value for
the coefficient of variation in both systems, as well as the minimum connectivity within these
systems when channel belts are compensationally stacked ($\kappa_{CV}$=1.0). In the unconfined lower
Wasatch Formation, a $\kappa_{CV}$ value of 0.33 would be needed for 100% connectivity between
channel belts. Conversely, if the channel belts are compensationally stacked with a $\kappa_{CV}$ of 1.0,
only 0.37% of channel belts are connected. In the confined Dakota Sandstone a $\kappa_{CV}$ value of 0.21
results in 100% connectivity, and fully compensational channel belts ($\kappa_{CV}$=1.0) have 2.52% of all
channel belts connected. Highly compensational channel belts have low connectivity because they are separated by overbank fines. Strongly clustered channel belts have high connectivity because they are in direct contact with one another with no overbank fines separating them.

Overall, we document that the cluster dimensions derived from point density maps, the R3 method, compensational stacking ($\kappa_{CV}$), and connectivity ($\gamma$ index) to be the most robust methods for documenting channel-belt stacking patterns in fluvial systems. Furthermore, the relationships between these metrics provide a conceptual framework for documented down-dip changes in channel-belt stacking patterns and the roles of confinement of fluvial systems.

2.7 Applications

The relationships between channel-belt clustering, compensation and connectivity are important in the evaluation of subsurface fluvial reservoirs. For example because $\kappa_{CV}$ is a surface based method, seismic data and correlations from well logs can be used to evaluate the variation. $\kappa_{CV}$ for the interval of interest can then be used along with an estimate of the degree of confinement for the system. Using these three variables in the power law connectivity equation (equation 10), the intra channel-belt connectivity can be calculated. This is important for predicting the connectivity between channel belts in sparse well environments, and for making well placement decisions during exploration and development. This is a direct application of the empirical results from the two end-member outcrops. Outcrops documented using $\kappa_{CV}$ for field specific applications would be best suited to characterizing the connectivity in the subsurface reservoirs, but the metrics are can be applied for both direct and indirect analogs assuming all other boundary conditions are similar.
The R3 method can be used in stochastic geostatistical reservoir simulations to seed the placement of channel belts, provided there is an appropriate outcrop analog. This inserts relationships from natural systems into the model rather than just placing channel belts from any number of random distributions.

2.8 Conclusion

Herein, we have quantitatively documented the following for an unconfined and a confined fluvial system: (1) confined systems have stronger channel-belt clustering, less compensational stacking of channel belts, and greater connectivity than their unconfined counterparts, (2) longitudinal changes in channel-belt clustering, compensational stacking, and connectivity, (3) cluster dimensions, and cluster shapes, and (4) two new workflows to predict and model fluvial channel belts at the basin scale.

This study uses four methods to document clustering, compensational stacking, and connectivity (Ripley’s K function, R3 Method, point density maps, $\kappa_{CV}$, and the $\gamma$ index) in both a confined (Dakota Sandstone) and unconfined (lower Wasatch Formation) fluvial system. These methods combined document the following: Ripley’s K function for clustering shows that unconfined and confined fluvial systems have different minimum clustering distances, and no major trend down dip in both systems. The R3 method documents that channel belts stack closer together in the confined system than in the unconfined system. Critically, channel belts stack closer together in the down dip outcrops than in the updip outcrops of the unconfined lower Wasatch Formation, while the confined Dakota Sandstone has consistent channel-belt stacking patterns from up-to-down dip. Next, from 3D channel belt density maps, we document that clusters become smaller down dip in both systems and that larger clusters have more channel belts by cross-sectional area. Using the coefficient of variation, we document that confined
systems have lower $\kappa_{CV}$ values than their unconfined counterparts, and that $\kappa_{CV}$ values decrease down dip in both the unconfined and confined systems. Furthermore, the $\gamma$ index documents that the confined Dakota has greater connectivity than the unconfined lower Wasatch System, and that inter channel-belt connectivity increases down dip in both the confined and unconfined systems.

Finally, we develop a conceptual framework for applying these new methods to reservoir prediction and modeling using the empirical relationship between connectivity and $\kappa_{CV}$. We then outline applications of the quantitative metrics for stochastic reservoir models in 3D using data directly from outcrop analogs.

2.9 Acknowledgements

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2.10 References


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Figure 2.1. (A) Satellite images of modern unconfined fluvial system (Gilbert River, Northern Australia) and cross sectional topography of the channel system. (B) Satellite images of modern confined fluvial system (Leaf River, Mississippi, USA), the transect across the valley documents the lateral boundary conditions that confine the channel belts within the valley. (C) Cross section of synthetic stratigraphy hypothesized for an unconfined system. (D) Cross section of synthetic stratigraphy hypothesized for a valley confined system.
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Figure 2.3. Depositional-strike oriented photographs of the three outcrops of the unconfined lower Wasatch Formation. Inset map shows the locations of the outcrops, which represent increasingly down-dip positions in the system. The locations of all channel belts are documented at each outcrop in order to test longitudinal changes in channel-belt clustering in an unconfined system.
Figure 2.4. Strike-oriented photographs of the three outcrops of the confined Dakota Sandstone. Inset map shows the locations of the outcrops in increasingly down-dip positions. The locations of all channel belts and the valley margin are documented at each outcrop in order to test longitudinal changes in channel-belt clustering in a confined fluvial system.
Figure 2.5. Diagram describing the R3 method for documenting clusters. (A) Cross section of a tank experiment showing channel belts (black) and their centroids (red) (Cross section from Paola and Martin, 2012). (B) Cross section showing centroids and the Manhattan distance component vectors between for two different reference points (for simplicity). Black lines correspond to reference point A, while blue lines correspond to reference point B. (C) Frequency histogram for all pairwise combinations of distances normalized by average channel widths (x) and thickness (y). The most common cluster has channel belts that stack 1.5 and 4 channel-belt thicknesses and widths from each other, respectively.
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Figure 2.9. (A) Channel-belt density cross sections for the updip, medial, and downdip outcrops of the unconfined lower Wasatch Formation. (B) Channel-belt density cross section for the updip, medial, and downdip outcrops of the confined Dakota Sandstone. (C) Transparent map view of channel belt density in the lower Wasatch Formation used to document minimum cluster lengths. (D) Transparent map view of channel belt density in the Dakota Sandstone used to document minimum cluster lengths.
Figure 2.10. Box and whisker plots of cluster widths (A), thicknesses (B), lengths (C), and aspect ratios (D). Clusters in the unconfined lower Wasatch are wider, thicker, and longer than those in the confined Dakota Sandstone. However, the aspect ratios of clusters in the confined Dakota Sandstone are greater than those in the unconfined lower Wasatch.
Figure 2.11. Log-log plot of number of channel belts and cluster cross-sectional area. As the clusters become larger in cross-sectional area the number of channel belts in the cluster increases following a linear trend for both systems. The unconfined lower Wasatch Formation and confined Dakota Sandstone both follow this trend, but there is no clear relationship between cluster area and longitudinal position in the basin.
Figure 2.12. Synthesis of differences between the confined (Dakota Sandstone) and unconfined (lower Wasatch Formation) fluvial systems. Clustering and connectivity is high in the confined (Dakota) system, while compensational stacking is low. Clustering and connectivity is low in the unconfined system (Wasatch) while compensational stacking is high. Longitudinally both systems document the same longitudinal trends. Cluster size decreases down dip, compensational stacking decreases down dip, and connectivity increases down dip.
Figure 2.13. Empirical relationships between the quantitative metrics used to document channel-belt stacking patterns. Connectivity between channel belts is strongly dependent on the compensation index ($k_{cv}$). Minimum clustering distance however, appears to have a weak to no quantitative relationship to connectivity, or the coefficient of variation.
Table 2.1. Table of the different documented characteristics in the confined (Dakota Sandstone) and unconfined (lower Wasatch Formation). Many of the key external boundary conditions are the same in both systems, documenting that lateral boundary conditions were the primary control on differences in stratigraphic architecture.

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<th>Characteristic</th>
<th>Unconfined lower Wasatch Formation</th>
<th>Confined Dakota Sandstone</th>
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<td>Primary Facies</td>
<td>Cross Stratified Sandstone</td>
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<tr>
<td>Primary Accretion Type</td>
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</tr>
<tr>
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</tr>
<tr>
<td>Proportion of crevasse splay</td>
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<td>Low</td>
</tr>
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<td>Average Thickness of Channel Belts</td>
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CHAPTER 3
ON THE RELATIONSHIP BETWEEN FLUVIAL CHANNEL-BELT MORPHOLOGY AND INTERNAL HETEROGENEITY: INSIGHTS FROM SATELLITE, SEISMIC, NUMERICAL, AND OUTCROP DATASETS

A paper to be submitted to The Journal of Geophysical Research: Earth Surface
J.R. Pisel, D.R. Pyles

3.1 Abstract

Despite the visible differences between meandering and braided rivers, little work has been done to relate plan-view morphometric measures of active channels to the channel belt deposited by the migration of rivers. Herein, we use satellite, outcrop, and published numerical and seismic datasets to investigate some process-based controls on channel-belt morphology and its relationship to active channel morphology. We find that channel-belt morphology is controlled primarily by the erodability of the substate, and the residence time of an active channel. Channels that migrate slowly with long residence times result in smoother channel belts than those that migrate rapidly with short residence times. Internally this means that channel belts with a high coefficient of variation (rugosity) in width are more likely to have more diverse facies proportions than channel belts with a low coefficient of variation. Despite this, identifying laterally accreting and downstream accreting channel belts based on their morphology remains difficult if the laterally accreting channel belt has had sufficient residence time on the floodplain. The control of lateral and subjacent lithology on channel-belt morphology is important for targeting subsurface reservoirs in 3D seismic data as well as predicting channel movement in modern-day river corridors.
3.2 Introduction

Fluvial channel-belt elements are the stratigraphic expression of the migration and evolution of rivers over geologic time scale (Ford and Pyles, 2014). Two end-member types of channel-belt elements are documented in modern and ancient fluvial systems, lateral and downstream accreting (Ford and Pyles, 2014). However, predicting facies heterogeneity within channel belts in the subsurface has remained challenging (Shepherd, 2009). Using modern rivers, published numerical studies and outcrop analogs has aided in predicting internal heterogeneity (Jordan and Pryor, 1992), but a quantitative relationship between the morphologies of modern rivers, outcrops, numerical simulations, and 3D seismic data remains elusive.

Herein, we present, for the first time, a predictive method for using external channel-belt morphology as a predictor of internal stratigraphic characteristics, such as active channel morphology. This study uses four complimentary data sets (satellite images of modern channel belts, published numerical simulations, seismic images of ancient channel belts, and exceptionally well exposed outcrops of ancient channel belts) in a process based framework for predicting channel morphology and ultimately the migration direction of modern fluvial channels.

3.3 Data

Data was compiled from four primary data sources. The modern rivers analyzed from satellite images span a range of latitudes from equatorial to arctic and tectonic regimes from passive margins to intracratonic basins. The published numerical and physical simulations have constrained sediment and water flux ratios. Seismic images are from a Pleistocene active rift basin located near the equator; whereas the channel belts in the outcrop studies were deposited...
during a greenhouse climate in a foreland basin. Spatially, the dataset spans multiple orders of magnitude from the physical experiments to the satellite images.

### 3.3.1 Modern Rivers – Satellite Images

In this study we selected 30 modern rivers (Table 3.1) using the following criteria. First, the active channel and channel belt must be mappable from the satellite image (Figure 3.1). Second, in multi-channel, anastomosed systems, the dominant channel is used rather than the other smaller channels (Figure 3.1). Finally, the datasets of the rivers must contain a full meander wavelength of the active channel. A range of sinuosity values documented in the dataset ensure that the results are applicable to all different types of rivers, and not just end-member examples.

From the satellite images the following data were collected (Table 3.2): (1) mapped lateral extents of the channel belts, (2) channel-belt widths ($W_b$), (3) active channel widths ($W_c$), and (4) orientation of abandoned channels and bar form accretion surfaces (Figure 3.1). From these remote sensed data, the following secondary data were calculated: (1) active channel sinuosity ($P$), (2) radius of curvature ($R_c$), (3) bend curvature ($R_c/W$), (4) wavelength ($L$), (5) amplitude ($A$), and (6) channel belt widths ($W_b$).

### 3.3.2. Ancient Rivers – Seismic Data

High resolution 3D seismic data were used to document map-view patterns of ancient fluvial channel belts and their associated abandoned channel fills, yielding a similar perspective to the satellite-derived data in that the planform morphology of the abandoned channel fill and channel belt are discernable from the associated floodplain belt (e.g. Figure 3.2a).

We used seismic time slices from a dataset collected in the Gulf of Thailand included in Samorn (2006). The seismic dataset has a frequency of ~70 Hz and the limit of detectability is ~6 m in thickness. The channel belts in the Gulf of Thailand dataset are Pleistocene to Holocene
age, and were deposited the extensional Pattani Basin. We investigated five channel belts in different time slices from this dataset. The channel belts were selected using similar criteria to the channel belts in the satellite dataset. In all five cases the channel belts were created by single-thread meandering rivers, and the channel abandonment fill records the final phase of the active channel before abandonment and avulsion to a new location on the floodplain. From the seismic time-slice images, the following data were collected: (1) mapped lateral and longitudinal extents of the channel belts, (2) channel belt widths, (3) abandoned channel fill widths ($W_c$), and (4) orientation of abandoned channels and bar form accretion surfaces (Figure 3.1; Table 3.2). From this data the following secondary data was calculated: (1) channel abandonment fill sinuosity ($P$) which is analogous to active channel sinuosity ($P$), (2) radius of curvature ($R_c$), (3) bend curvature ($R_c/W$), (4) wavelength ($L$), (5) amplitude ($A$), and (6) channel belt widths ($W_b$).

### 3.3.3. Ancient Rivers – Outcrop data

The second set of ancient fluvial channel belts documented in this study, outcrops, provide a similar perspective as seismic data in that the planform morphology of the channel belts was documented rather than the cross-sectional stratigraphy. However, due to modern erosion only portions of the extents of the channel belts are preserved. Three world-class outcrops in the Morrison Formation, Cedar Mountain Formation, and Dakota Sandstone were chosen for this study because of the three-dimensional exposures of both laterally and downstream-accreting channel belts (Figures 3.3, 3.4). The channel belts contain three-dimensional exposures of bar accretion surfaces and sedimentary structures (Figure 3.4). The spatial location of the paleocurrent indicators such as cross-stratification, flutes, ripples, and channel belt margins constrain the morphometric measurements of the channels in each outcrop and the locations of sand dominated barforms.
The Jurassic Morrison Formation of eastern Utah is 180-200 m thick and composed of the Tidwell, Salt Wash, and Brushy Basin Members (Figure 3.3) (Lupton, 1914; Gregory, 1938; Peterson, 1988; Kjemperud et al., 2008). The Morrison Formation unconformably overlies the Jurassic Summerville Formation, and is unconformably overlain by the Neocomian through Albian age Cedar Mountain Formation (Currie, 1998; Kjemperud et al., 2008). The Morrison Formation was deposited in the back-bulge region of the foreland basin created by the Nevadan Orogeny to the west (DeCelles and Currie, 1996; Currie, 1998). The Salt Wash Member is a high net-sand content fluvial system composed of channel belt and floodplain-belt elements (Craig et al., 1955). Laterally and downstream-accreting channel-belt elements of the Salt Wash Member are exceptionally well exposed in three-dimensions south-east of Green River, Utah (Figure 3.4) and are a part of this study. Paleocurrent measurements document the Salt Wash Member channel belts flowed north to north-east, transverse across the basin away from the thrust belt to the east towards the forebulge located to the west (Figures 3.3) (Currie, 1998).

The Lower Cretaceous Cedar Mountain Formation of eastern Utah is 75 m thick in eastern Utah and is subdivided into 5 different Members, (1) Buckhorn Conglomerate, (2) Yellow Cat Member, (3) Poison Strip Member, (4) Ruby Ranch Member, and (5) Mussentuchit Member (Figure 3.3) (Garrison et al., 2007). The Cedar Mountain Formation unconformably overlies the Jurassic Morrison Formation (Currie, 1998). In eastern Utah the Cedar Mountain is unconformably overlain by the Mowry shale while in central Utah it is unconformably overlain by the Cenomanian Dakota Sandstone (Cobban 2007; Kirschbaum and Schenk, 2010). The Cedar Mountain Formation was deposited in both the backbulge and forebulge of the Nevadan-Sevier basin (Currie, 1997). All members of the Cedar Mountain formation have been documented as either fluvial or lacustrine (Garrison et al., 2007). Both laterally and downstream-
accreting channel-belt elements of the Ruby Ranch Member outcrop in three-dimensions southwest of Green River, Utah and are a key part of this study (Figure 3.4). The Cenomanian Dakota Sandstone of central Utah is 40 m thick and was deposited near the crest of the forebulge of the fully formed Sevier basin during the Late Cenomanian. It unconformably overlies the Cedar Mountain Formation, and is conformably overlain by the Tununk Member of the Mancos Shale (Figure 3.3) (Kirschbaum and Schenk, 2002). The Dakota Sandstone study area has been interpreted by Kirschbaum and Schenk (2010) to be an exceptionally preserved fluvial system confined within a valley. The preservation of dominantly fluvial channel belts is unique as most incised valley systems are typically filled with estuarine deposits (Kirschbaum and Schenk, 2010). Channel belts are dominantly downstream accreting in the lower Dakota, while the upper portion has both downstream and laterally accreting channel belts (Kirschbaum, pers. comm.). Both styles of channel-belt elements within the Dakota are exhumed so that they are accessible in three-dimensions (Figure 3.4). Paleocurrent measurements document channel belts of the Dakota Sandstone flowed northeast, longitudinally parallel to the Sevier thrust belt to the east (Kirschbaum and Schenk, 2010).

Within the Salt Wash Member of the Morrison Formation, Ruby Ranch Member of the Cedar Mountain Formation, and the Dakota Sandstone six exhumed channel-belt elements in total are documented in this study, two in each formation (Figures 3.4). We document the following characteristics: (1) strike orientation of bar forms, (2) dip orientations of ripples, parting lineations, cross strata foresets and troughs, (3) laser range finding measurements of channel-belt widths, and (4) abandoned channel fill widths (W). From this primary data secondary data includes: (1) channel abandonment fill sinuosity (P) which is analogous to active
channel sinuosity, (2) radius of curvature \(R_c\), (3) bend curvature \(R_c/W\), (4) wavelength \(L\), (5) amplitude \(A\), and (6) channel belt widths \(W_b\) (Table 3.2).

### 3.3.4 Published Numerical Simulations

Fifty-seven numerical and physical simulations from published datasets were used in this study, following the same criteria for the modern rivers documented from satellite images (Van Dijk et al., 2012; Asahi et al., 2013; Nicholas et al., 2013; Nicholas, 2013a; Nicholas, 2013b). The 57 simulations contain only straight, low and moderate sinuosity rivers (Table 3.2). The numerical simulations have a mixture of sediment types and vegetation in both the channel belt and floodplain belt (Asahi et al., 2013; Nicholas et al., 2013; Nicholas, 2013a; Nicholas, 2013b).

From the simulations the following data were collected: (1) mapped lateral and longitudinal extents of the channel belts, (2) channel belt widths, and (3) active channel widths \(W_c\) (Figure 3.1). From this data the following secondary data was calculated: (1) active channel sinuosity \(P\), (2) radius of curvature \(R_c\), (3) bend curvature \(R_c/W\), (4) wavelength \(L\), (5) amplitude \(A\), and (6) channel belt widths \(W_b\) (Table 3.2).

### 3.4 Methods

In this study we document both morphology of the channel belt and the morphology of the active channel or final position of active channel when abandoned in all datasets and develop a new method for quantifying channel-belt morphology using the coefficient of variation of channel-belt widths. Furthermore, we measure active channel morphology using traditionally used metrics such as channel width \(W_c\), sinuosity \(P\), radius of curvature \(R_c\), bend curvature \(B_c\), wavelength \(L\), and amplitude \(A\) (Figure 3.1).
3.4.1 Channel-Belt Morphology

Fluvial channel-belt elements are the stratigraphic expression of the migration and evolution of rivers over geologic time scale (Ford and Pyles, 2014) (Figure 3.1). Measures of channel-belt morphology or "rugosity" have been proposed by Payenberg et al. (2014). The different proposed measures use either perimeter line lengths of the channel belts, or centerline perpendicular channel-belt width measurements. The measures fail to quantitatively differentiate between downstream and laterally accreting channel belts because downstream and laterally-accreting channel belts have similar rugosity values. Building on this approach, we developed an alternate measure of channel belt rugosity using the coefficient of variation ($CV_R$) which is defined as:

$$CV_R = \frac{\sigma_w}{\bar{x}_w}$$

where $\sigma_w$ is the standard deviation of numerous widths from a single channel belt along a downstream transect, and $\bar{x}_w$ is the mean of the population. In this measure, as $\sigma_w$ of channel belt width approaches zero, $CV_R$ approaches zero ($\lim_{\sigma_w \rightarrow 0} CV_R = 0$). Also, as $\sigma_w$ exceeds $W_b$, $CV_R$ goes to 1 ($\lim_{\sigma_w \rightarrow W_b} CV_R \geq 1$).

We approach measuring channel-belt widths in a different manner than has been formerly used (e.g. Payenberg et al., 2014), which is to measure width along a path normal to the channel-belt centerline. The issues with that approach are the following (Figure 3.5): (1) in highly rugose channel belts the width measures often overlap which in turn biases the measure; and (2) the areas where there is more variation in channel belt width are sampled at the same resolution as the areas with less variation, potentially over emphasizing the straight areas in relation to the more complex areas (Figure 3.5). To address these issues, we used a width calculation approach
that adapts to the local morphology of the channel belt. In this method we first use Voronoi polygons to partition the channel belt within a convex hull (Figure 3.5). Voronoi polygons contain one point within each cell, and the boundaries between cells are equidistant from the two adjacent points at all lengths (Voronoi, 1908). We define points on the margin of the channel belt as the seed points for each Voronoi polygon. Once polygons are calculated for each channel belt, we clip them within the channel belt polygon, meaning the generated Voronoi polygons define the area within the channel belt (Figure 3.5). Next, we convert the perimeters of the Voronoi polygons to lines (Figure 3.5). Then we calculate the standard deviation ($\bar{\sigma}_w$) and mean width of the channel belt ($\bar{x}_w$) from the Voronoi perimeter lines for the entire channel belt, which is, in turn, used to calculate $CV_r$ (Equation 1) for each channel belt.

This approach is much more robust than previously used methods as the width measures of the Voronoi polygons are highly sensitive to areas along the margin of the channel belt that are highly complex in shape.

3.4.2 Active Channel Morphology

In contrast to the ancient rivers, where only channel belt morphology and abandoned channel fills can be documented, the numerical and satellite datasets have channels that are transporting water and sediment. The active channels are used herein to quantitatively relate sediment-transport processes to channel-belt morphology, and ultimately to predict lithology of the strata located the subjacent and lateral to the channel belt.

Channel width ($W_c$) is measured perpendicular to the active channel centerline from inner to outer bank along the main channel (Figure 3.1). Sinuosity ($P$) is defined as the distance along the channel centerline divided by the straight line distance for one meander wavelength of the channel, and is one of the most common measures used to compare modern rivers (Figure 3.1).
Radius of curvature \((R_c)\), is defined as the distance from the channel inflection point to the bend apex for one meander wavelength (Figure 3.1). Bend curvature \((B_c)\) is calculated by dividing the radius of curvature by the width of the active channel (Figure 3.1) and is nondimensional.

Wavelength \((L)\) is measured from one bend apex to the corresponding apex after two inflection points along the channel centerline (Figure 3.1). Amplitude \((A)\) is defined as the perpendicular distance from one bend apex across the channel belt to the opposing bend apex (Figure 3.1). We use these measures to document the morphology of the active channel for all datasets, however in the seismic and outcrop datasets we used a proxy for the active channel.

For the seismic dataset, we measure sinuosity using the abandoned channel fill to approximate active channel morphology at the time of abandonment. For the outcrop dataset we used methods proposed by Le Roux (1991; 1994) to calculate sinuosity for the active channel during deposition. This method uses the operational range of paleocurrent orientations (Le Roux, 1994). Sinuosity is calculated using the following set of equations:

\[
P = \pi \left( \frac{\phi}{360} \right) / \sin \left( \frac{\phi}{2} \right)
\]

(2)

when the operational range \((\phi)\) is less than 180° and:

\[
P = \pi \left( \frac{\phi}{360} \right) / \sin \left[ \frac{(360-\phi)}{2} \right]
\]

(3)

when the operational range is greater than 180°. The operational range \((\phi)\) is defined as 3.2 times the circular standard deviation of the paleocurrent measurements. This calculates an accurate measure of sinuosity \((P)\) given a large paleocurrent dataset from ancient channel belts.

The active channel measurements from the satellite, numerical, and seismic datasets were then cross-plotted against \(CV_R\) to investigate relationships between channel-belt morphology and active channel sediment transport processes that can be inferred from active channel morphology (Figure 3.5). We then investigate how process-based controls influence the stratigraphic
architecture of channel belts and infer internal heterogeneity from qualitative observations from the outcrop dataset.

### 3.5 Results

Measures of channel-belt and active-channel morphology span several orders of magnitude in dimensional space, and vary significantly between datasets (Table 3.2). Rugosity and sinuosity have significantly less variance than the other morphometric measures as they are normalized domains, but have significant spread nonetheless (Table 3.2).

Rugosity ($CV_R$) values for the channel belts from the satellite dataset range from 0.55 to 1.21 (Table 3.2), while rugosity values for the numerical simulations range from 0.29 to 0.85 (Table 3.2). The seismic dataset has rugosity values from 0.59 to 0.61. Number of active channels ($N$) for all datasets range from 1 to 9 (Table 3.2). Sinuosity ($P$) values vary from 1.04 to 2.69 for the satellite dataset, and 1.01 to 1.78 for the numerical simulations (Table 3.2). The seismic dataset has sinuosities from 1.86 to 2.55, and the outcrop dataset has values of 1.11 to 1.73 (Table 3.2). Bend curvature ($R_c/W$) varies from 1.75 to 14.1 for all datasets, while active channel widths ($W_c$) range from 8 m to 8,008 m (Table 3.2).

The plots in Figure 3.6 document little-to-no relationships between rugosity ($CV_R$) and active channel measurements. The cross-plot of rugosity ($CV_R$) as a function of sinuosity ($P$) (Figure 3.6A) shows the most robust relationship between channel-belt morphology and active-channel morphology. At low sinuosity ($P$) values, rugosity ($CV_R$) increases abruptly before reaching a steady state at a value of approximately 0.7. There is no clear correlation between channel sinuosity and internal heterogeneity in the datasets. The channel belts with scroll bars do not have any significantly higher sinuosity values than the sand prone braided channel belts (Figure 3.4).
Because empirical stratigraphic relationships between channel-belt morphology and channel sinuosity are not evident, below we evaluate process-based controls on channel-belt morphology, and to this end we build upon the seminal work of Parker (1976), Ikeda et al (1981) and Sun et al. (1996). In their article, Sun et al. (1996) document the erosion coefficients of floodplain deposits and channel-belt deposits to be primary controls on preserved channel-belt morphology. When the erosion coefficient of the floodplain deposits is smaller than that of the underlying channel belt deposits the channel preferentially erodes through the underlying channel-belt deposits creating a smooth channel belt with low $CV_R$. In order for this to happen, the channel must translate downstream one full meander wavelength (Figure 3.8). Based on this concept, we interpret channel belt rugosity to be dependent on the downstream migration rates $\left( \frac{\partial x}{\partial t} \right)$, lateral migration rates of the active channel $\left( \frac{\partial y}{\partial t} \right)$, wavelength ($L$) and amplitude ($A$) of the channel. This interpretation is based on the assumption that an active channel with long wavelength, high amplitude, high lateral migration rates, and low downstream migration rates has a channel belt with larger $CV_R$ value than a similar channel with low lateral migration rates and high downstream migration rates. This assumption is a starting point expressed as:

$$CV_R \sim \frac{A}{L} \left( \frac{\partial x}{\partial t} \right) \left( \frac{\partial y}{\partial t} \right)$$  \hspace{1cm} (4)

If the assumption that lateral and downstream accretion rates are controlled primarily by the erosion coefficient of the floodplain using equation 19 from Sun et al. (1996), our equation 4 simplifies to:

$$CV_R \sim \frac{AEX}{LEY}$$  \hspace{1cm} (5)
where $E_x$ is the cross-stream erosion coefficient, and $E_y$ is the downstream erosion coefficient.

Furthermore, we define a nondimensional smoothing time scale ($T_{s*}$) to quantitatively document the time for channel-belt rugosity to reach 0:

$$T_{s*} = \frac{A}{\frac{E_x}{L} \frac{E_y}{E_x}}$$  \hspace{1cm} (6)

or:

$$T_{s*} = \frac{A E_y}{L E_x}$$ \hspace{1cm} (7)

where values of $T_{s*}$ greater than one document that channel-belt rugosity will never approach 0 and values less than one document that channel belt rugosity will decrease with decreasing $T_{s*}$ (Figure 3.7).

Next we use avulsion frequency ($f_A$) defined by Jerolmack and Mohrig (2007, their equation 1):

$$f_A = \frac{v_a N}{\bar{h}}$$ \hspace{1cm} (8)

where $v_a$ is the in channel aggradation rate, $N$ is the number of active channels and $\bar{h}$ is the channel depth. To calculate a timescale over which channel belts reach low rugosity ($CV_R$), we divide $T_{s*}$ by avulsion frequency:

$$T_R = \frac{T_{s*}}{f_A}$$ \hspace{1cm} (9)

where $T_R$ is the time for channel belt rugosity ($CV_R$) to approach 0. Substituting the right hand sides of equations 7 and 8 we document the rugosity timescale as:

$$T_R = \left(\frac{A E_y}{E_x L} \frac{\bar{h}}{v_a N}\right)$$ \hspace{1cm} (10)

which relates rugosity ($CV_R$) to amplitude ($A$), wavelength ($L$), depth ($\bar{h}$), number of active channels ($N$), and aggradation rate in the active channel ($v_a$) in addition to the lateral ($E_x$) and
downstream accretion rates ($E_y$). Low $T_R$ values predict channel belts reach low $CV_R$ values faster than those with high $T_R$ values.

We then combine $T_R$ with the avulsion timescale of Jerolmack and Mohrig (2007) to calculate what we herein term the roughness index ($R$), which refers to the rugosity ($CV_R$) of a channel belt given its residence time on the floodplain before abandonment. The avulsion timescale is defined as:

$$T_A = \frac{\bar{h}}{v_a}$$  \hspace{1cm} (11)

where $v_a$ is the in-channel aggradation rate and $\bar{h}$ is the channel depth. The ratio between the rugosity timescale ($T_R$) and the avulsion timescale ($T_A$) is:

$$R = \frac{T_R}{T_A}$$  \hspace{1cm} (12)

when the rugosity index ($R$) is greater than 1 ($R > 1$) the active channel will avulse to a new location on the floodplain before channel-belt rugosity ($CV_R$) decreases, or smooths out (Figure 3.8). Whereas, where $R < 1$ the active channels will be in place long enough for channel belts to reach low rugosity ($CV_R$) values (Figure 3.8). Substituting the right hand sides of equations 10 and 11 into the right hand side of equation 12 gives:

$$R = \frac{AE_y}{NLE_x}$$  \hspace{1cm} (13)

documenting that the rugosity index depends on the amplitude, wavelength, number, downstream and lateral accretion rates of the active channel (Figure 3.8). Equation 13 is rearranged to solve for the ratio of lateral accretion to downstream accretion rates:

$$\frac{E_x}{E_y} = \frac{A}{NLR}$$  \hspace{1cm} (14)

which documents that the ratio of lateral-to-downstream accretion rates can be calculated given amplitude ($A$), number of channels ($N$), wavelength ($L$), and channel belt rugosity ($CV_R$).
To test the efficacy of this model we use the modern Arkansas River near Rosedale, Mississippi (Figure 3.9a). Using satellite images from 1984 to 2012 we document part of the evolution of the channel belt. We find that the measured $CV_R$ is less than that predicted by equation 4 (Figure 3.9b). We attribute this to the variance in meander amplitudes and wavelengths along the reach of the channel, along with the error in calculating migration rates from satellite images with 1 meter resolution.

Using Equation 6 we calculated a mean $T_{s^*}$ of 0.19 (Figure 3.9c), meaning the channel belt will decrease in rugosity with time. Indeed, Figure 3.9a documents that the measured rugosity of the channel belt does decrease as a function of time. Note that with only 1 active channel $T_{s^*}$ and the rugosity index ($R$) are the same. We then assumed sedimentation rates were constant at 0.3 cm/yr (Leeder, 1978) and used equation 8 to calculate an avulsion frequency of $3.29 \times 10^{-4}$ (1/years). We then combined our results from equations 6 and 8 to calculate a mean $T_R$ of 599 years (Figure 3.9d). Using the same sedimentation rate assumption we calculated an avulsion timescale of 3,033 years (Equation 11) which is significantly longer than the mean rugosity timescale of 599 years, meaning the channel belt will effectively smooth itself out before the channel avulses to a new location on the floodplain. We then investigated the large difference between predicted and measured $CV_R$ from 2005 to 2010. We find that during this time the ratio of wavelength to amplitude decreased, even though the ratio of down-to-cross stream migration rates were roughly constant (Figure 3.9e). This change in wavelength to amplitude ratio means that the amplitude is increasing relative to the meander wavelength, and increasing the predicted rugosity in Figure 3.9b, but providing an explanation to the discrepancy between measured and predicted $CV_R$. 

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3.6 Discussion

Despite our understanding of process-based controls on the evolution of channel-belt morphology, there is still considerable uncertainty in differentiating between channel-belt types and styles using seismic channel-belt morphology. Multistory, laterally-accreting channel-belt elements in the seismic dataset have similar rugosity values as multistory, downstream-accreting channel-belt elements from the satellite dataset (Figure 3.6); meaning that despite the quantitatively documented differences in internal architecture and facies, the overall channel-belt morphology are similar when viewed, for example in planview 3D seismic (Figures 3.2, 3.6). However, single-story, laterally accreting channel belts and multi-story, downstream accreting channel belts have different rugosity, which in turn provide a basis for differentiating between the two in the subsurface (Figure 3.10). Single-story channel belts migrating laterally in one direction, relative to their active channel inflection points, have higher rugosity values than multi-story channel belts migrating laterally in both directions relative to their active channel inflection points. The migration in more than one direction causes the channel belt to smooth out over time (Equation 13, Figure 3.8B), while migration in one direction causes the channel belt to become more rugose because the area between meander loops is not filled with sand during later periods of migration, as we hypothesize in the multi-story channel belts.

From a process-based perspective, channel-belt morphology is important for understanding the composition of both the channel belt and underlying floodplain. In the case where channel belts have high sinuosity but low rugosity (Figure 3.8B) this implies that the channel belt material has higher erosion coefficients than the floodplain (Equation 14), and can be used to infer sand-rich channel belts for both laterally and downstream accreting systems. Conversely, in systems with high rugosity and high sinuosity (Figure 3.8c) it is most likely that
although the channel belt may be sand rich, the adjacent floodplain is much more mud rich and easier to erode, with a high lateral to downstream migration rate (Equation 14). Therefore channel-belt rugosity can be used as a predictor of the subjacent and lateral lithology. Other potential variations on highly rugose channel belts can stem from early carbonate cementation of sands within the channel belt which increases the erosion coefficient, making it easier for the active channel to erode into the floodplain. In the outcrop dataset (Figure 3.4A (Right)), within the Dakota Sandstone Kirschbaum and Schenk (2010) qualitatively document the systems with higher downstream to lateral accretion rates to have higher proportions of mud preserved within the channel belt. This in turn is most likely linked to fine grained preservation potential within the channel belts. Therefore systems with higher lateral accretion rates are less likely to remobilize previously deposited sediment within channel belts than systems with less lateral accretion (Equation 4).

Allogenic controls, such as slope, discharge, confinement, sediment supply, and grain size have all been interpreted to influence channel morphology. Herein we agree with previous studies that bank cohesion and erosion are a dominant control on not only active channel morphology (Parker, 1976; Millar, 2000), but we also attribute these controls to the resulting channel-belt morphology. We have shown that allogenic controls such as uplift and subsidence rates set the timescale over which external controls dominate, but over autogenic timescales, such as the rugosity timescales, the organization of the transport system dominates. Therefore, the roughness index (Equation 12) documents the ratio of autogenic timescales and places quantitative limits on primary autogenic controls within the sediment routing system and their stratigraphic expression.
For modern systems, Equation 14 is important for predicting the future migration paths of active channels. This in turn provides insight into both the composition of the underlying floodplain erosion in both cross and downstream directions. Furthermore, this ratio can be used to calculate channel motion and predict where the river will migrate. This is important for engineering modern river corridors along major population centers. Additionally, this pattern can be used to predict lithologies in subsurface reservoirs (Figure 3.9).

3.7 Conclusion

Herein, we quantitatively documented that sinuosity is most closely related to channel-belt morphology. However, the relationship is based more on physical processes occurring within the channel and adjacent floodplain, rather than on an empirical relationship between the two morphometric measures. We propose that channel-belt rugosity is directly related to measureable characteristics of the active channel, along with physical properties of the floodplain. Furthermore, we present a conceptual framework that uses the rugosity timescale and avulsion timescale of a river system to calculate the roughness index of individual channel belts and document the tradeoff between autogenic and allogenic timescales. This in turn can be used to predict the shapes and migration directions of both modern and ancient channel belts given set parameters and timescales. Further experimental, numerical, and modern river field work is needed to validate this conceptual framework and explore the active subspaces of this model.

3.8 Appendix of Variables

$W_c$ active channel width

$P$ active channel sinuosity

$R_c$ radius of curvature

$R_c/W$ bend curvature
L  meander wavelength
A  active channel amplitude
\( \bar{\sigma}_w \)  standard deviation of channel belt width
\( \bar{x}_w \)  mean channel belt width
CV\(_R\)  coefficient of variation – rugosity
Φ  operational range of paleocurrent data
\( \frac{\partial x}{\partial t} \)  active channel lateral accretion rate with respect to time
\( \frac{\partial y}{\partial t} \)  active channel downstream accretion rate with respect to time
E\(_x\)  cross-stream erosion coefficient
E\(_y\)  downstream erosion coefficient
T\(_{s*}\)  smoothing timescale
f\(_A\)  avulsion frequency
ν\(_a\)  channel aggradation rate
N  number of active channels
\( \bar{h} \)  channel depth
T\(_R\)  rugosity timescale
T\(_A\)  avulsion timescale
R  roughness index

3.9 References


Figure 3.1. (A) Schematic diagram of the different morphometric measures for the active channel and channel belt used herein. (B) Satellite image of a modern fluvial channel (active channel) and channel belt of the Rio Negro in Argentia as an example of how systems were documented in this study (Image from Google Earth, 2015).
Figure 3.2 (A) Selected images from seismic and modern datasets. Seismic amplitude time slice of a fluvial channel belt from the Gulf of Thailand (from Samorn, 2006). The channel belt, abandoned channel, and point bars are visible in the seismic data. The channel belt has high rugosity and the abandoned channel has high sinuosity (B) Satellite image of the Missouri River in the United States with lower rugosity channel belt and high sinuosity active channel. (C) Satellite image of the Milk River in Canada with higher rugosity channel belt and high sinuosity active channel.
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Figure 3.3. (A) Chronostratigraphic chart for the Late Jurassic through Middle Cretaceous (After Curry, 1997). (B) Map of Utah showing the locations of outcrops documented in this study relative to the Sevier highlands during the early Cretaceous. The study intervals are highlighted by the gray rectangles.
Figure 3.4. Satellite images of the outcrop channel belts outlined in yellow with geospatially referenced and oriented paleocurrent indicators as blue arrows and accretion surfaces as strike and dip symbols. All channel belts are shown at the same scale for the (A) Dakota Sandstone, (B) Cedar Mountain Formation, and (C) Morrison Formation. Additionally, rose diagrams are plotted next to each channel belt and document the mean orientations and dispersal patterns.
Figure 3.5. Satellite images displaying the different methods for documenting channel-belt widths. (A) Measuring channel belt widths perpendicular to the channel-belt centerline produces gaps and overlaps in the areas measured, whereas Voronoi polygons measure at consistent spacing throughout the channel belt. (B) Annotated satellite images documenting the steps used to construct the Voronoi polygons and polylines used to measure channel belt width. See text for specifics.
Figure 3.6. Cross plots of channel-belt rugosity ($CV_R$) as a function of the active channel morphometric measurements from Figure 3.1. Sinuosity vs. rugosity (A) document the most consistent trend while the other measures (B, C, D) have little-to-no relationships between channel-belt morphology and active channel morphology.
Figure 3.7. Three-dimensional plots of the nondimensional smoothing timescale ($T_s^*$) as a function of active channel amplitude ($A$) and wavelength ($L$). (A) Ratio of cross-stream to downstream erosion coefficients less than 1, and (B) ratio of cross-stream to downstream erosion coefficients greater than 1. Channel belts that fall below the $T_s^*$=1 plane will decrease in rugosity than those that plot above the plane. Low $T_s^*$ values document shorter nondimensional time scales for decreasing rugosity.
Figure 3.8 Diagram showing synthetic channel belts and the morphological differences due to cross and downstream migration. (A) A synthetic channel belt translating downstream where the avulsion timescale ($T_A$) is shorter than the rugosity timescale ($T_R$). (B) Synthetic channel belt translating downstream where the avulsion timescale ($T_A$) is approximately the same as the rugosity timescale ($T_R$). (C) Synthetic channel belt documenting that increased wavelength ($L$) of the active channel means the channel must translate further downstream to decrease channel belt rugosity. (D) Synthetic channel belt with only lateral migration documenting increasing rugosity ($CV_R$) with high values of $T_s^*$ and a high roughness index ($R$).
Figure 3.9. (A) Satellite image of the Arkansas and Mississippi Rivers in 2012 and inset map with location of the rivers. (B) Time series plot of predicted and measured rugosity for the Arkansas River in (A). (C) Time series plot of the smoothing timescale calculated for the Arkansas River, note increase in 2005 that corresponds to the decrease in wavelength to amplitude ratio in (E). (D) Time series plot of rugosity timescale for the Arkansas River with an increase in the number of years for the channel belt rugosity to decrease that corresponds to the same wavelength to amplitude ratio decrease in (E). (E) Time series plot comparing the down-to-cross stream migration rate and wavelength to amplitude ratio of the Arkansas River in (A).
Figure 3.10. Satellite images and channel belt morphology after 30 years of channel belt evolution for (A) laterally-accreting channel belt; (B) a laterally and downstream-accreting channel belt, and (C) a downstream-accreting channel belt. The upper row are satellite images of modern rivers while the lower row is the channel belt morphology after 30 years of active channel migration. The lower channel depicts how the channel belt would appear in plan view at reservoir depths in 3D seismic.
Table 3.1. Table of the names, latitude, and longitude of the rivers included in the satellite image dataset. The dataset spans 110 degrees of latitude and 323 degrees of longitude.

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Table 3.2. Comparison of rugosity, sinuosity, number of active channels, radius of curvature, channel width, and bend curvature for the four different datasets used in this study. Values include the minimum, maximum, mean, and median calculated for each dataset. Satellite data has the largest variance in all the measures.

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CHAPTER 4
QUANTITATIVE ANALYSIS OF THE FLUVIAL STRATIGRAPHIC FILTER AT THE
CHANNEL BELT SCALE: LOWER WASATCH FORMATION, UTAH, USA

A paper to be submitted to Geology
J.R. Pisel, D.R. Pyles

4.1 Abstract

This article uses measurements from five fluvial channel belts of the Paleocene lower Wasatch Formation to quantitatively document the transience or persistence of flow velocities recorded in stratigraphy at the bedset scale. We use facies proportions and sedimentary structures coupled with a paleomorphodynamics workflow to calculate the mean flow velocity for each bedset. Flow velocity measurements were analyzed using a lattice approach that documents either persistence or transience of mean flow velocities, which, in turn was combined with facies trends to infer perennial and ephemeral flow conditions interpreted to have been caused by allogenic fluctuations. Three of the five channel belts have significant spatial dependence of mean flow velocities. Based on short-range spatial dependence, we infer perennial flow conditions in a laterally-accreting channel belt, and ephemeral flow conditions in two downstream-accreting channel belts. The remaining two channel belts have no spatial dependence as the stratigraphic filter has completely destroyed the allogenic signal within the channel belts. Furthermore, we document that intra channel-belt signal preservation comes at the expense of basin-scale signal preservation, meaning high frequency, allogenic signals effectively erase low frequency allogenic signals from the stratigraphic record.
4.2 Introduction

External forcing (allogenic) mechanisms such as changes in tectonic uplift, subsidence rates, solar insolation (Milankovich cycles) and climate fluctuations have been documented in fluvial stratigraphy and simulated in forward numerical models (Foreman et al., 2012; Allen et al., 2013; Allen et al., 2014). Time series methods have been used to document allogenic signals, including spectral analysis of Fourier and wavelet transforms, as well as autocorrelation functions (Prokoph and Agterberg, 1999; Prokoph and Bilali, 2008; Jerolmack and Paola, 2009). However, recent research documents that internal (autogenic) mechanisms act as a non-linear filter that can destroy allogenic signals if the amplitude or period of the allogenic signal is less than the autogenic morphodynamic turbulence of the system (Jerolmack and Paola, 2009). Recognizing allogenic signals in stratigraphy is important for predicting how fluvial systems respond to tectonic, orbital, and climatic changes. Allogenic signals have been documented at the basin scale (100-1,000 m scale thickness), however few studies have concentrated on the channel-belt and bar scale (1-100 m thick) (e.g. Allen et al., 2014).

Two end-member channel-belt types have been interpreted to document short term (yearly) climatic signals, perennial and ephemeral. Perennial channel belts are interpreted to document persistent flow conditions, with annual fluctuations related to seasonality in sediment and water flux (Meinzer, 1923; Fielding et al., 2009). Ephemeral channel belts document transient flow conditions between wet and dry periods (McKee et al., 1967; North and Taylor, 1996). Despite differences in flow conditions and bedforms associated with the two channel belt types, documenting persistence or transience of flow conditions within ancient fluvial channel belts remains challenging. This study uses cross sections, measured sections, grain-size
distributions, lithofacies, and bedset bounding surfaces to document allogenic climate signals within fluvial channel belts.

4.3 Geologic Setting

The lower Wasatch Formation of the Uinta Basin in eastern Utah contains world class exposures of a low net-sand content fluvial system. The Uinta Basin is a longitudinally asymmetric foreland basin located in northeastern and central Utah (Figure 4.1a). From the Paleocene through Eocene, flexurally induced subsidence provided accommodation for deposition of the Green River and Wasatch Formations (Figure 4.1a). Lacustrine sediments deposited in the center of the basin were surrounded by deltaic and fluvial strata of the Green River and Wasatch Formations; signifying internal drainage (Picard, 1955; Keighley et al., 2002). Paleocurrent directions in the southern outcrops of the Wasatch Formation document fluvial systems flowing north and northeast towards the center of the basin (Ford and Pyles, 2014) (see paleocurrent rose diagram in Figure 4.1b). Climatically, the lower Wasatch Formation is interpreted to have been deposited during global hot house conditions (Wilf, 2000). Basin-scale studies in the adjacent Piceance basin document million-year changes in channel belt dimensions and sedimentary structures attributed to climatic fluctuations at the Paleocene-Eocene Thermal Maximum (Foreman et al., 2012).

4.4 Dataset and Methods

An exceptionally well exposed, strike-oriented outcrop of the lower Wasatch Formation is used to address the goals of this study (Figure 4.1b). The outcrop is located along the southern margin of the Uinta basin, just west of the modern day Green River, and is 5 km wide by 300 m thick. The outcrop contains 274 fluvial channel belts, all of which are exceptionally well exposed and accessible. Five channel belts were analyzed in detail. They represent the range of
architectural variability in the outcrop and span a range of varying numbers of stories and accretion styles. Using the hierarchical approach of Ford and Pyles (2014) the 5 channel belts were characterized on the basis of bar migration direction as follows (Figure 4.2): (Channel Belt 1) downstream-accreting single story, (Channel Belt 2) downstream-accreting multi story, (Channel Belt 3) laterally-accreting with erosionally based fine-grained fill multi story, (Channel Belt 4) laterally-accreting multi story, (Channel Belt 5) downstream and laterally accreting multi story, respectively (Figure 4.2).

The following data were collected to address the goals of this study: (1) decimeter-resolution measured sections that qualitatively documents grain-size distributions, sorting, rounding, physical and biogenic sedimentary structures, bedset, story, and element boundaries; (2) high-resolution photo panels; (3) paleocurrent orientations collected from flutes, ripples, cross-strata, channel-belt margin orientations; and (4) laser range finding measurements of element, story, and bar form widths and thicknesses. These data were used to generate further information about the channel belts using the following workflow. First, grain size distributions were calculated from measured sections. Median grain size ($D_{50}$) and maximum grain size ($D_{95}$) are calculated from the distributions for each channel belt (Figure 4.3a). Cross-sections of the channel belts were created by tracing bedset boundaries in the photo panels and combined with measured sections to constrain grain size and facies type for each bedset. Next, using paleoslope reconstruction methods of Lynds et al. (2014), the mean slope was calculated from all 5 channel belts. The average slope for the lower Wasatch Formation is $3.01 \times 10^{-3}$. Fourth, measured bar-form heights were used to constrain flow depths for bedsets (Figure 4.3b). Finally, paleoslope, median grain size, and flow depth measurements (Figure 4.3c) were used with the Law of the Wall to calculate mean flow velocity for each bedset.
Spatial persistence and transience of mean flow velocity were quantified using spatial
statistics and lattice methods. Lattice data are discrete, with each region represented by an
average of the data. In this study we defined regions by bedsets, and assign mean-flow velocity
to each. We defined spatial neighborhoods for each bedset using row standardized weights,
meaning that bedsets in contact with one another (linked) are spatially related. Beyond adjacent
bedsets, we evaluated spatial autocorrelation of mean flow velocity at increasing, non-adjacent
bedset lags using Moran’s $I$. Bedset lag spatial autocorrelation simply increases the
neighborhood structure to include beds that are not directly in contact with one another which
documents long-range spatial trends. Spatial autocorrelation, which is the cross-correlation of a
region with its neighbors is calculated using Moran’s $I$:

$$I = \frac{\sum_{i=1}^{n} \sum_{j=1}^{n} w_{ij} (y_i - \bar{y})(y_j - \bar{y})}{\sum_{i=1}^{n} \sum_{j=1}^{n} w_{ij} \sum_{i=1}^{n} (y_i - \bar{y})^2}$$

(1)

where $y_i$ is the $i$-th observation, $y_j$ is the $j$-th observation $\bar{y}$ is the global mean flow velocity and
$W_{ij}$ is the spatial weight of the link between regions $i$ and $j$ defined above using row standardized
binary weights (Moran, 1950). The expected value of Moran’s $I$ under the null hypothesis of no
spatial dependence is:

$$E(I) = \frac{-1}{N - 1}$$

(2)

where $N$ is the number of locations. To test for spatial autocorrelation using Moran’s $I$, we use
Monte Carlo simulations. In this test, 99 Monte Carlo simulations were run for each channel belt.
Values for each region are randomly reassigned to a new region and Moran’s $I$ is calculated for
each simulation. Calculated Moran’s $I$ is compared to the distribution of Moran’s $I$ from the
Monte Carlo simulations. If the observed value of $I$ is outside the distribution generated from the
simulations (p<0.05), there is significant evidence for spatial autocorrelation. Moran’s $I$ ranges from -1 to 1, where negative values document negative correlation, and positive values document positive correlation. Values near 0 document spatial independence.

4.5 Results

For a bedset lag of 1, Channel Belts 3 and 4 have documented positive spatial autocorrelation (or similarity) of mean flow velocity using Moran's $I$ (Figure 4.4). However, increasing region neighborhoods, or distance between bedsets, we document positive spatial autocorrelation of only Channel Belt 3 up to 2 bedset lags (Figure 4.4). Meaning there is short-range positive spatial dependence of mean flow velocities in Channel Belts 3 and 4. In contrast, Channel Belt 1 has negative spatial autocorrelation of mean flow velocity (Figure 4.4). This means there is short-range negative spatial dependence of mean flow velocities. Channel Belts 2 and 5 have no correlation for any bedset lags (Figure 4.4) meaning that mean flow velocities for all bedsets are spatially independent or different.

In all 5 channel belts the diagnostic sedimentary structures associated with high and low flow regimes coupled with facies proportions provide further information into the type of signal preserved. Specifically, spatial persistence of flow velocity, low facies diversity, and low flow regime associated facies are interpreted to be characteristic of perennial deposits. In contrast, spatial transience of flow velocity, high facies diversity, and high flow regime associated facies are interpreted to be characteristic of ephemeral deposits.

We interpret the short-range, positive autocorrelation, facies, and bar migration orientation in Channel Belts 3 and 4 to collectively record preserved allogenic signals. Both of these channel belts migrated solely laterally, which is interpreted to have fully preserved the depositional processes compared to channel belts that migrated in the downstream direction,
which results in partly eroded bedsets. Furthermore, using facies proportions, we interpret Channel Belt 3 to document ephemeral deposits as the facies record high flow regime (e.g. facies F8, F9, F10) and vary significantly within the channel belt. Channel Belt 4 is interpreted to document perennial deposits as the facies are predominantly low flow regime associated facies (e.g. facies F3, F4, F5) and have little variability within the channel belt. Therefore, we infer that the allogenic climate signal was preserved by the lateral accreting barforms, as the potential for erosion of the bedsets after deposition decreases significantly as the active channel continued to migrate in the lateral direction, away from older bedsets.

The negative spatial autocorrelation, facies, and bar migration orientations in Channel Belt 1 collectively document a preserved climatic signal that is diagnostic of ephemeral flow conditions. Although Channel Belt 1 is a downstream accreting channel belt, the negative autocorrelation documents transience (or difference) in mean flow velocity at one bedset lag. Large fluctuations in flow velocity are characteristic of ephemeral rivers where there are periods of low to no flow followed by periods of high flow. A majority of facies in Channel Belt 1 (e.g. Facies F9 and F10) are characteristic of high deposition rates that are common in ephemeral deposits, and support the hypothesis of a preserved climate signal (Figure 4.4). Furthermore, the variability of facies which reflect alternating flow velocities within the channel belt support the interpretation of ephemeral deposition within Channel Belt 1. We interpret that the fluctuations between wet and dry were large enough in period or amplitude to record times of both increased and decreased mean flow velocity.

The complete spatial independence of mean flow velocities, facies, and migration direction in Channel Belts 2 and 5 are collectively used to interpret completely destroyed allogenic signals. Despite the spatial independence of bedsets in these two channel belts, we
interpret Channel Belt 2 documents perennial deposits; the majority of the facies are diagnostic of lower flow regime conditions which are interpreted by North and Taylor (1996) to record low discharge and low flow velocity conditions (Figure 4.4). Facies in Channel Belt 5 document both high and low flow conditions, but is primarily facies that are interpreted to document low flow regime (Figure 4.4). Therefore, we interpret this channel belt to be a combination of perennial and ephemeral flow conditions, with no spatial signal. Also, both these channel belts are dominated by downstream accretion, which means there is lower preservation potential.

4.6 Discussion

From these results, we interpret the rate of lateral migration within laterally-accreting channel belts, to be proportional to preservation potential of the subjacent bars. Results from this study indicate intra-channel belt signal preservation to be opposite those of the basin-scale patterns documented by Straub and Esposito (2013). At the bar scale, if a channel moves laterally quickly the underlying basin-scale strata is removed, but the intra-channel belt architecture is preserved. Conversely, if a channel doesn’t migrate laterally, the basin-scale architecture is better preserved, but the intra-channel belt architecture is removed due to downstream migration, effectively shredding signals via the stratigraphic filter. Therefore there is a scale-dependent tradeoff in signal preservation from intra-channel belt to basin-scale architecture. This concept provides insight into the scales that future studies should consider when attempting to resolve allogenic signals. Systems with deep laterally-migrating channel belts should be considered ideal when attempting to resolve signals at the intra-channel belt scale, while systems with shallow downstream-migrating channel belts should be considered ideal to study when attempting to resolve basin-scale signals.
4.7 Conclusion

We interpret perennial and ephemeral fluvial systems in the lower Wasatch Formation based on spatial dependence and facies types. Perennial rivers have short range positive autocorrelation, and are composed of low-flow regime bedforms. Ephemeral rivers have both short range positive and negative autocorrelation, no long range spatial dependence, and contain sedimentary structures and facies indicative of upper flow regime and high deposition rates. Signals that are completely masked by the stratigraphic filter document no spatial dependence at all distances. However, facies proportions and sedimentary structures document facies associated with both perennial and ephemeral rivers.

This article quantitatively documents, for the first time, preserved and destroyed allogenic climate signals within fluvial channel belts. We use a paleomorphodynamic workflow to calculate mean flow velocity for each bedset, and using Moran’s $I$, facies patterns, and migration orientations, we document spatial dependence and independence in mean flow velocity. We interpret short-range spatial dependence and facies inconsistency to document a preserved signal, and complete spatial independence and facies consistency to document autogenic shredding of the intra-channel belt signal. Concepts developed in this study provide context from a world class fluvial outcrop to previous tank and numerical studies on the autogenic “shredding” of allogenic signals, and are applicable to both modern and ancient fluvial systems around the globe.

4.8 Acknowledgements

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Scholarship, and the Hemmesch and Burch Fellowships from the Colorado School of Mines. We thank Duncan Metcalfe, Corinne Springer, Dave Potter, and the Natural History Museum of Utah for field support and access to Desolation Canyon.

4.9 References


Figure 4.1. (a) Map showing the location of the Uinta Basin and bounding structures. Inset is a chronostratigraphic chart of the Uinta Basin. The lower Wasatch Formation lies between the Flagstaff Limestone and middle Wasatch Formation and is the focus of this study (See Supplemental Figure A). (b) Photopanel of the field area showing the locations of the 274 channel belts. The five used in this study are labeled 1-5 and are highlighted by the colored boxes (See Supplemental Figure A).

Figure 4.2. Stratigraphic cross sections through the five channel belts within the lower Wasatch Formation (See Supplemental Figure A for more detail). Sediment transport is into the page for all channel belts. Bedsets are colored according to the dominant facies within the bedset. Brown and green colors reflect facies with clay-sized sediment, while yellow and orange are facies with silt and sand. Facies proportions by area are documented in the pie charts next to each channel belt.
Figure 4.3. (a) Grain-size distributions for the channel belts used in this study. Grain-size measurements were made throughout the measured sections, and across the outcrop face. (b) Box and whisker plots of flow depths measured from bar-form thicknesses within the channel belts. (c) Box and whisker plots of mean flow velocities calculated for bedsets in channel belts.

Figure 4.4. Spatially lagged Moran’s I values for channel belts in the lower Wasatch Formation. Increasing the neighborhood structure to include bedsets not directly in contact with one another documents both short and long-range changes in spatial dependence.
CHAPTER 5

CONCLUSIONS TO DISSERTATION

As seen in the opening paragraph of this dissertation, this chapter is the conclusions to the dissertation. It is organized in a similar manner to Chapter 1. I discuss the summary of conclusions and scientific contributions of each individual chapter. I start with Chapter 2, followed by Chapter 3, and conclude with Chapter 4.

5.1 Summary of Conclusions and Contributions

This dissertation uses outcrop datasets from the Paleocene lower Wasatch Formation, Cenomanian Dakota Sandstone, Albian Cedar Mountain Formation, and Tithonian Morrison Formation coupled with remote sensing data from satellite imagery and 3D seismic data to investigate autogenic and allogenic controls on fluvial systems at multiple scales. These scales range from decimeter to 10’s of kilometers in space, and from minutes to millions of years in duration (Figure 5.1). Specifically, this dissertation progresses our scientific understanding of how allogenic controls are fundamentally replaced by either the mechanics of sediment transport or preservation potential.

5.2 Conclusions and Contributions of Chapter 2

At the largest temporal, spatial and process scale (Figure 5.1), Chapter 2 documents how lateral boundary conditions influence channel belt stacking patterns and inter channel-belt connectivity in 3D. Furthermore this chapter tests and develops new quantitative metrics to document how fluvial channel belts are spatially distributed in stratigraphy. Specifically, in the valley-confined Dakota Sandstone, the lateral boundary conditions cause channel belts to stack more closely together and are therefore better connected than those in the unconfined lower Wasatch Formation. Additionally, the channel belts in both the confined and unconfined systems
become more strongly clustered in the down-dip outcrops. At the basin scale (10’s to 100’s of km), allogenic boundary conditions are directly transferred into the preserved stratigraphy. One application of these concepts is to constrain hydrocarbon reservoir models.

5.3 Conclusions and Contributions of Chapter 3

Moving to the intermediate temporal, spatial and process scale (Figure 5.1), Chapter 3 documents how fluvial channel-belt morphology is directly controlled by a combination of allogenic and autogenic controls. This scale is characterized by time spans of 100’s to 1000’s of years temporally and 0.1 to 10,000 km spatially. While the external controls set how long a channel belt is in one location, this dissertation quantitatively documents that the dynamics of the sediment routing system ultimately determine the resulting morphology of the channel belt. Furthermore, it is difficult to quantitatively relate active channel morphology and channel-belt morphology. Ultimately this chapter documents that although channel belts can share similar morphology given enough residence time, their accretion type and internal architectures can be vastly different. Therefore, caution should be used when predicting accretion type of an active channel based on channel-belt morphology when targeting these reservoirs using 3D seismic.

5.4 Conclusions and Contributions of Chapter 4

At the smallest temporal, spatial and process scale (Figure 5.1), Chapter 4 quantitatively documents for the first time how allogenic signals can be detected and preserved. This chapter focuses on the deposition of bedforms over minutes to days temporally and decimeters to 10’s of meters spatially. Chapter 4 develops a paleomorphodynamic workflow from previously published work to calculate the mean flow velocity at the time of deposition of each bedform and documents the spatial relationship of flow velocity within each channel belt.
This chapter documents that in order for signals to be preserved in stratigraphy, the rate of deposition must either be very high or the active channel must be migrating rapidly so as to not remobilize the previously deposited sediment. Furthermore, the internal architecture and facies proportions of channel belts with preserved allogenic signals are statistically different than channel belts dominated by autogenic controls. Therefore, when modeling intra channel-belt architecture for hydrocarbon reservoirs, the stratigraphic expression of allogenic and autogenic controls must be considered for an accurate model.
Figure 5.1. Synthesis diagram for Chapters 2-4 of this dissertation. (A) Time scale from seconds to millions of years, (B) spatial scales from microns to 100’s of kilometers, (C) dominant processes at different time and space scales. Chapter 2 focuses on allogenic controls at large spatial and temporal scales, while Chapter 3 investigates autogenic controls at intermediate space and time scales, and Chapter 4 is concerned with grain-to-grain interaction at the micron spatial scale and seconds time scale.
Appendix A. Chapter 4 supplemental figure with location map, depositional strike oriented photo, and stratigraphic cross sections used to document allogenic signals in fluvial channel belts.
APPENDIX B

Satellite Images of Rivers – SUPPLEMENTAL ELECTRONIC MATERIAL

Appendix B includes satellite images of the 30 modern rivers documented in Chapter 3. In this file, the satellite images are displayed at the same scale and numbered in order that they are listed in Table 3.1.

| B-1_Satellite_River_Images.PDF | Satellite images of 30 modern rivers |
APPENDIX C

Permission Letters – SUPPLEMENTAL ELECTRONIC MATERIAL

Appendix C includes copies of letters from co-authors of the journal articles associated with Chapters 2, 3, and 4. In these letters, the coauthors grant permission for the use of the articles as chapters in this dissertation.

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