Measuring channel mobility through the analysis of area-based change in analog experiments, with insights into alluvial environments

by

Andrew D. Wickert

Submitted to the Department of Earth, Atmospheric and Planetary Sciences in Partial Fulfillment of the Requirements for the Degree of Bachelor of Science in Earth, Atmospheric, and Planetary Sciences at the Massachusetts Institute of Technology

31 May 2007

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Abstract

We quantify the mobility of alluvial river channels in order to better understand the relationship between channel mobility and fluvial environment. We analyze five analog experiments, performed between 2002 and 2007 at the Saint Anthony Falls Laboratory, for area-based change in river-channel plan form with time. These experiments isolate the effects of (1) sediment input and deposition, (2) base-level, and (3) bank cohesion on channel mobility.

(1) In experimental deltaic environments with aggradation and sea-level rise in equilibrium, aggradation rates scale non-linearly with channel mobility. The channels in the experiment with the higher aggradation rate move more rapidly overall, but the channels in the experiment with slower aggradation are more mobile when both experiments are scaled by their respective aggradation rates. Lower aggradation rates result in lower slopes, causing the flow to be shallower and broader, and for more deposition to occur. These low slopes result in lower Froude number, allowing the formation of flow-depth-high ripples that divert the flow. (2) We study experiments in which base-level change was slow, rapid, and intermediate in terms of average channel mobility when base-level is constant. Channels in an experimental basin with rapid (12.2 mm/hr) aggradation are three times as mobile during base-level rise as during base-level fall. Channels in a cohesive delta experiment (producing significantly lower channel mobility) are 1.5 times as mobile during slow base-level rise (0.237 mm/hr) as during steady base level. Base-level change that is slow on an average channel mobility timescale has little to no influence on channel mobility. (3) Channels in an experimental system of noncohesive sediment are 6.5 times as mobile as channels with alfalfa (Medicago sativa) planted in the flume.

Thesis Supervisor: John Southard
Title: Professor of Geology Emeritus
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All of the SAFL personnel who help to run the laboratory and the experiments, as this work wouldn’t be possible without them.

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Chapter 1: Introduction

Alluvial river channels migrate and avulse, interacting with their environments to shape the depositional record and affect the life around them. They exist at an interesting confluence, where the forces of nature in the solid, liquid, and gaseous earth meet. This confluence is biological as well: Riparian ecosystems change the ways in which channels move, often restricting their form and slowing their ability to migrate by surrounding them with vegetation and its mechanical effects on soil. In the very recent geologic past, human activities have also come to shape river channels through great works of agriculture and civil engineering. These many interactions make river channels complex, and inspire scientists to discover the incompletely-understood physical reasons for why and how they evolve.

Our goal in this study is to investigate one of the primary large-scale characteristics of alluvial river channels: that they can move. Rivers migrate laterally and avulse by abandoning their old channels and establishing new ones. By moving in different ways, they create plan forms that are as varied as the sinuous Mississippi, the anastomosed Orinoco delta, braided glacial rivers, and the complex patterns of streams that form on alluvial fans (Jerolmack and Mohrig, 2007). Their motions also develop topography; build the stratigraphic record; and bring food, water, infrastructure, and recreation to people and civilizations.

The rates at which channels are mobile are incompletely understood. Therefore, we seek to create and refine a new method for measuring river channel change with time and
Figure 1.1. **Williams River in Alaska.**

(above right) Fully meandering. (Photo by N. D. Smith.) **Figure 1.2. Orinoco Delta.**

(above left) Delta Amacuro, Venezuela. (NASA GeoCover LANDSAT) **Figure 1.3**

(below) **Alluvial fan with braided channels.** Svalbard, Norway. (Photo by Rich Miller)

to apply it to understanding rates of channel motion on experimental systems. In particular, we investigate the effects of deposition (through sediment input and base-level changes) and bank stability. The advantage of experimental systems is that we are able
control the boundary conditions and sample extremely dense data sets. Therefore, they provide the proper resources to test the development of a new method, and to try to discern simple relationships between channel mobility and fluvial environment.

1.1. Scientific Background and Our Work

Changes in channel plan form occur by avulsion (abandonment of a former channel in favor of a new course) and lateral migration (side-to-side motions caused by erosion and deposition on river banks). Which of these occurs, and the rate at which these occur, are dependent on a number of properties of a channel system. These properties are channel pattern (i.e. single-thread or multi-thread; wide and shallow or deep and narrow), bank stability (which is related to bank material, vegetation, and bioturbation of the bank), water discharge (which is related to drainage-basin-wide meteorology and climate, and the speed at which precipitation is transported into a drainage network), and sediment flux (including erosional and depositional effects that aggrade or degrade the bed).

Bryant et al. (1995) and Heller and Paola (1996) show that system-wide aggradation rate is the primary control on avulsion frequency in experimental settings. They demonstrate that avulsions occur, on average, when a channel has aggraded a full channel depth, leaving the base of a channel perched atop its floodplain. This condition is termed a superelevation of 1, as the channel has aggraded one full depth. Their findings were supplemented with a field study of channel deposits in outcrop by Mohrig et al. (2000). Their findings were consistent with the experiments, and showed that the rock record also contains evidence of avulsions occurring on average at a superelevation of 1.
Channel migration is related to the plan form and sinuosity of the channel, which are in turn determined by its environment. The work of Parker (1976) and Jerolmack and Mohrig (2007) show that specific channel patterns form in response to fluid dynamical and sedimentological instabilities. These are correlated to channel migration rate because of the relationship of these instabilities with specific channel geometries. In our work, however, we look only at the processes forming the plan forms, and not at the plan forms themselves, as the main causative agents of channel mobility. This is consistent with Parker (1976) and Jerolmack and Mohrig (2007), who show that channel plan form can be modeled as a result of channel-forming processes that can be defined through other intrinsic variables, such as sediment flux, water flux, and slope. So while there is a correlation between channel mobility and river morphology, we feel confident in stating this correlation as a relationship between channel mobility and the factors responsible for shaping river morphology, and thus removing it from our analysis. (We are additionally confident in this decision because we are averaging over a very large amount of data, such that individual morphological effects (e.g., the shape of a particular element of sinuosity in a channel at a particular time) should be averaged out of the final results.)

Increased bank stability slows channel mobility (especially migration) by reducing erosion rates. The result of increasing bank stability on plan form is to cause the channel to change from a braided pattern to a sinuous pattern with a lower braiding index and deeper, narrower channels. This can be caused by the encroachment of vegetation, which has been described by experimental studies (e.g., Smith, 1976; Gran and Paola, 2000;
Simon and Collison, 2001; Tal et. al., 2004; Tal and Paola, 2007) and field work (e.g., Simon and Collison, 2001; Micheli et al., 2004). Riparian vegetation increases stream-bank stability through root reinforcement, rain interception, and the production of cohesive, biota-rich soils. Increased bank stability can also be caused by increased bank cohesion (e.g., Hoyal and Sheets, manuscript in preparation, 2007), in that noncohesive sediment can be eroded and redeposited more easily than a cohesive substrate. These effects are only partially understood insofar as they provide limitations on the rates of channel migration due to the material properties of the environment of the channel because these material properties are variable and the result of complex natural systems. Examples of this variation are different sedimentary layers (with different material properties), dynamic vegetative and ecosystem responses, and the variability and lensing nature of deposits within certain depositional environments, such as those formed of glacial debris. Whatever the mechanism, increased bank stability causes the formation of deeper, narrower channels that move more slowly than the threads of an unconstrained braided stream.

Water flux is important in determining channel mobility, because it is responsible for erosion and sediment transport. Water flux depends on drainage-basin climate, meteorology, size, shape, and surface properties (i.e., speed of water delivery). The most important relationship between water flux and channel mobility is the recurrence interval of bankfull floods, because they are responsible for shaping channels (e.g., Wolman and Miller, 1960). Because we wished to avoid these complexities of drainage-basin hydrology and flood recurrence intervals, the experiments we performed held flow
consistently at bankfull to slightly overbank discharge in all but one case (in which a recurrence interval was applied in order to allow simulated vegetation time to grow). All of our analysis was during times of bankfull discharge, so we considered only channel-shaping flows. We were then able to use the relations of Parker et al. (in press, 2007) to determine that channel volume should scale with water flux. However, we decided to compare only experiments with the same water flux in order to constrain our variables more rigidly.

Deposition plays a crucial role in determining the mobility of alluvial channels. As already stated, aggradation within a channel causes superelevation and avulsion (Bryant et al., 1995; Heller and Paola, 1996; Mohrig et al., 2000). Jerolmack and Mohrig (2007) linked channel migration to deposition by turning the superelevation criterion on its side, applying it to lateral rather than vertical movement. The result is the relationship that a channel should either have migrated one channel width or have avulsed when it has deposited a full channel volume of sediment. (Of course, the migration of one channel width is in a spatially averaged sense: in a meandering case, for example, the most active areas of the bends might migrate several channel widths while other parts of the channel migrate an extremely short distance.) Therefore, we believe that channel mobility should scale with the deposition – either on the bed or on banks. One interesting case of this that we seek to understand is the effect of changes in base level on channel mobility. Changes in base-level change the locations of erosion and deposition on the delta surface. Base-level fall results in incision and entrenchment of channels and progradation of a long-lived, stable deltaic lobe. Base-level rise results in zones of aggradation moving inland as
the shoreline retreats. This decreases deltaic-lobe time scale and causes main channels to fill and avulse more frequently.

1.2. An Overview of Our Work

We approach this study with three major goals. First, we seek to develop a highly generalizable method for monitoring change in fluvial pattern on a fluvially active surface (Chapters 1-4). Second, we seek to apply this method to a series of experiments, in which the boundary conditions are tightly controlled but autogenic processes are allowed to function as usual, albeit at a smaller scale (Chapters 5-6). Third, we seek to use our measured values for channel mobility in these experiments, with respect to their experimental boundary conditions, to draw relationships between the environments and channel mobility (Chapter 7).

The work presented in this thesis is a combination of thousands of hours spent running experiments and analyzing data. The experimental methods were developed and the experiments were performed by Sheets et al. (in press, 2007), Martin et al. (manuscript in preparation, 2007), Kim (2006; 2007), Hoyal and Sheets (manuscript in preparation, 2007), and Tal and Paola (2007). The images were processed and prepared for analysis by Ben Sheets, John Martin, Wonsuck Kim, Michal Tal, and the author. The analysis was developed by the author with help from Wonsuck Kim and Matt Wolinsky. All of the analyses were performed by the author, except for the 2007 Cohesive Sediment Experiment, in which John Martin did the greater part of the work.
Chapter 2: Basic Methods
(A Stepwise Introduction to the Decorrelation Analysis)

2.1. Introduction

We analyzed a series of time-lapse images of experimental fluvial systems by separating wet (mostly channelized) pixels from dry (bar, bank, or floodplain) pixels, and then by comparing the locations of these pixels throughout a time series of photos. In order to perform this analysis, we colored the water in these fluvial experiments with a dye (either blue #9 or rhodamine) as a proxy for depth. This dyed water flowed over a mobile substrate with a different color, and therefore could be differentiated from the rest of the system by digitally separating the colors. This analysis allows us a basic glimpse at the mobility of the channels within the experimental system.

The following subsections illustrate the steps taken to (1) change the images into binary matrices of wet and dry pixels, and (2) compare these wet-dry matrices with one another in order to which quantify the dissimilarity that exists between two images in terms of their in wet and dry area, discretized into pixels. Chapters 3 and 4 explain preliminary results, observations, and the more sophisticated (and often more insightful) methods of analysis that were developed in order to supplement the basic methods for comparison developed in this section, and to understand how plots of decorrelation relate to channel mobility.
2.2. Selection

Some sets of time-lapse images were taken on a time scale much shorter than the time scale in which channels are noticeably mobile. In order to analyze these sets of images while conserving computing time, they were coarsened such that noticeable channel movement (migration and/or avulsion) existed between one image and the next.

Images were also selected by quality. In sets of images with inconsistent lighting and/or dye concentrations, a set of images in which these factors were reasonably consistent was hand-selected for analysis. In addition, images which contained anomalous elements (graduate students, professors, SAFL personnel, pieces of equipment, other obstructions, and the shadows of all of those listed) were discarded from the analysis.

2.3. Ortho-Rectification, Image Stitching, and Masking

Each of the selected photographs of the channel system was adjusted to provide a view from above at an angle orthogonal to the surface dip. This ortho-rectification was performed using the panorama tools plug-in developed independently for use with Adobe Photoshop.

In order to observe channel mobility on an experimental deltaic surface properly, it was necessary to limit our analysis to the subaerial surface of the delta. This was in order to observe the proper proportions of water and land (“wet” and “dry” pixels) and to not allow the analysis to directly “see” the migration of the shoreline in the same way as it would “see” the migration of the channels. Therefore, the areas outside of the
Figure 2.1. Ortho-rectification of DB03-1. The left photo is an example of an original image taken during the experiment. The ortho-rectified image, which projects the view perpendicular to the surface dip angle of the delta, is on the right. These photos are of the high-resolution stratigraphy experiment (Sheets et al., 2007, in press).

Experimental tank and past the coastline of the delta were masked for the duration of the analysis.

In the case of the Riparian Vegetation and Braided Stream Dynamics experiment, which was performed in a flume that measured 2 meters by 16 meters, one camera could not adequately image the whole experimental surface. Therefore, four cameras stationed at approximately equal intervals above the flume were used to record morphological changes over time. After the images were taken, they were first individually rectified to correct the angles and remove lens distortions and then stitched together to form a composite panorama of the experimental channel network at each 30-second time step (Tal and Paola, 2007).
2.4. Separating “Wet” and “Dry” pixels

In each of these experiments, the water was filled with dye (blue #9 or rhodamine) to separate it visually from its substrate (quartz sand, anthracite coal, kaolinite, cohesive mixed sediment, and/or alfalfa). By using the color difference between the dye and the sediment and the increasing concentration of the color of the dye with flow depth, the picture of the experiment (top image in fig. 2.2) was transformed into a color-based “depth map” (middle image in figure 2.2). We thresholded this depth map in order to create a final binary matrix where each entry represented an individual pixel on the image. In this matrix:

- 1 is water (“wet”)
- 0 is land (“dry”)

Once we created binary wet-dry matrix, we refocused the object of the analysis toward understanding the relationships between the wet and dry pixels over a specified time series, in order to observe changes in channel location and their corresponding rates.
2.5. Notation for the Time Steps

The series of experimental images, transformed into binary wet-dry matrices, can be defined to include \( k \) experimental time steps (i.e. \( k \) binary wet-dry matrices), indexed \( K_l \) for \( l = 1, 2, 3, \ldots, k \). The best way for visualizing these time steps is on a number line, as in Table 2.1.

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2.6. Instantaneous Decorrelation

We define “instantaneous decorrelation” to be the bulk number of pixels that change between two subsequent binary wet-dry matrices, normalized by the amount of time in between the two images that the matrices represent. This analysis is designed to find two separate factors within a fluvial system. First, the average number of pixels that change per unit time between the two time steps closest to one another can be discerned, and this can be used as a measure of change within the fluvial system. Second, individual spikes and dips of the value of instantaneous decorrelation indicate times at which the channel system is either staying in place or is extremely active in migration or avulsion.

In order to measure the instantaneous decorrelation in the system, each binary matrix of wet and dry pixels is subtracted from the subsequent binary wet-dry matrix. The
subtraction of $K_{i+1} - K_j$ yields a matrix in which each element describes the change in its corresponding pixel as follows:

- 0 means “no change”
- 1 means “land changed to channel” OR “dry changed to wet”
- -1 means “channel changed to land” OR “wet changed to dry”

In order to simplify the system and look only at the total number of decorrelated pixels, we take the absolute value of this matrix. The sum across all of the entries of the new matrix gives the bulk instantaneous decorrelation ($D_{\text{inst}}$) over the period of time ($\Delta t_{\text{inst}}$) between the two images (time-steps) being analyzed.

$$\Delta t_{\text{inst}} \cdot D_{\text{inst}} = \sum_{i=1}^{m} \sum_{j=1}^{n} |K_{i+1} - K_j|$$  \hspace{1cm} (2.1)

In Equation 2.1, $m$ is the number of rows in the matrix, $i$ signifies any particular row, $n$ is the columns of rows in the matrix, and $j$ signifies any particular column.

In order to frame this value properly in terms of time (instead of in terms of arbitrary time steps), it is necessary to normalize by the amount of time ($\Delta t_{\text{inst}}$) between the two subsequent time steps. This normalization is especially for cases in which the time steps $l$ are not even, and $\Delta t_{\text{inst}}$ is therefore not constant. Although decorrelation with time follows an exponential decay curve (see Sections 4.2 and 4.3 for more detail on this), it can be approximated as linear for short $\Delta t$. The amount of time between two images is short, so this scaling to $\Delta t_{\text{inst}}$ works as a decent approximation for the instantaneous decorrelation analysis.
When plotted with $D_{\text{inst}}$ on the $y$ axis and time from $t_1$ to $t_k$ on the $x$ axis, the resultant graph shows time-normalized instantaneous decorrelation for the duration of the experiment. The graphical form is a good way to visualize both the average values of instantaneous decorrelation, as well as events like avulsions, flow expansions, and periods of large-scale erosion and deposition.

This method is not heavily used in this thesis but is planned to be developed further in subsequent work. One of its great assets is its ability to find avulsions and analyze short-time-scale events in channel systems. However, one of its limitations is that the amount of change between the first and second images is unnaturally high, and we hypothesize that this is due to noise in our analytical technique, where minor fluctuations in color and lighting cause larger-than-expected changes. A way to eliminate this noise is needed before this method can accurately display the proper amount of change between two images. One solution we propose is to look at the change between $K_{t+1} - K_t$ and $K_{t+2} - K_{t+1}$, so we ignore the drastic amount of change that occurs between the first and second images. We plan on examining this proposed solution in future work.
2.7. Decorrelation of a Time Series

2.7.1. Introduction

In order to understand how channels are mobile through time, we decided to compare a time series of wet-dry matrices against a selected baseline wet-dry matrix. It is expected that, for a baseline image $K_B$, there would be a little difference between the wet-dry matrices for $K_B$ and $K_{B+1}$. A slightly greater difference would exist between $K_B$ and $K_{B+2}$, and so on, until the plan forms of the channel systems are completely unrelated. This is indeed the case, and is the means by which we develop a method for determining a decorrelation time scale, which ultimately leads to a measure of river-channel mobility.

2.7.2. Methods

For a set of $k$ images as described in Section 2.2.4, individual binary wet-dry matrices $K_I$ exist for $I = 1, 2, 3, ..., k$. In order to discover how a specific channel system loses memory of its initial shape with time, a baseline wet-dry matrix ($K_B$) is chosen, and "transient" matrices $K_T$ such that $T = [B, B+1, B+2, ... k]$, are subtracted in turn from $B$. The equation for one of these subtractions is almost identical to the equation for non-time-averaged instantaneous decorrelation:

$$D_{\text{series}} = \sum_{i=1}^{m} \sum_{j=1}^{n} |K_T - K_B|$$  \hspace{1cm} (2.3)
In Equation 2.3, $D_{\text{series}}$ is the bulk decorrelation over a time series of transient wet-dry matrices with respect to a baseline wet-dry matrix. We calculate this subtraction for each transient matrix ($K_T = K_B, K_{B+1}, K_{B+2}, \ldots, K_k$), as shown in Table 2.2. “B” means that this time step is chosen as the baseline, and “T” means that the calculation was performed for this transient matrix. The term $t_i$ gives the run time of each particular time step.

The result is a series of data that shows the time steps analyzed, the runtime at which this time step occurred, the baseline matrix chosen, and the amount of decorrelation between the chosen baseline and each of the successive matrices. A chart of this is given in Table 2.3, with $t_i$ giving the run time of each particular time step.
Table 2.2. Performing the analyses with baseline and transient image-matrices.

<table>
<thead>
<tr>
<th>Time Step</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
<th>10</th>
<th>...</th>
<th>...</th>
<th>...</th>
<th>( k )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Time</td>
<td>( t_1 )</td>
<td>( t_2 )</td>
<td>( t_3 )</td>
<td>( t_4 )</td>
<td>( t_5 )</td>
<td>( t_6 )</td>
<td>( t_7 )</td>
<td>( t_8 )</td>
<td>( t_9 )</td>
<td>( t_{10} )</td>
<td>( t_{11} )</td>
<td>( t_{12} )</td>
<td>( t_{13} )</td>
<td>...</td>
</tr>
<tr>
<td>Steps in Analysis</td>
<td>BT</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
</tr>
<tr>
<td>B = baseline chosen</td>
<td>BT</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
</tr>
<tr>
<td>Analysis performed with</td>
<td>BT</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
</tr>
<tr>
<td>this image being the transient matrix</td>
<td>BT</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
<td>T</td>
</tr>
</tbody>
</table>

This table shows time step, runtime, and which analyses are performed. “B” indicates the selected baseline matrix, and T indicates the transient matrices for which the analysis is performed, with respect to the baseline matrix in the corresponding row.

Table 2.3. Calculating decorrelation with time

<table>
<thead>
<tr>
<th>( D_{\text{series}}(K_B) )</th>
<th>( t_1 \rightarrow t_2 )</th>
<th>( t_3 )</th>
<th>( t_4 )</th>
<th>( ... )</th>
<th>( t_k )</th>
</tr>
</thead>
<tbody>
<tr>
<td>( D_{\text{series}}(K_B=K_1) )</td>
<td>( K_B - K_B =0 )</td>
<td>( K_{B+1} - K_B )</td>
<td>( K_{B+2} - K_B )</td>
<td>( K_{B+3} - K_B )</td>
<td>( ... )</td>
</tr>
<tr>
<td>( D_{\text{series}}(K_B=K_2) )</td>
<td>-</td>
<td>( K_B - K_B =0 )</td>
<td>( K_{B+1} - K_B )</td>
<td>( K_{B+2} - K_B )</td>
<td>( ... )</td>
</tr>
<tr>
<td>( D_{\text{series}}(K_B=K_3) )</td>
<td>-</td>
<td>-</td>
<td>( K_B - K_B =0 )</td>
<td>( K_{B+1} - K_B )</td>
<td>( ... )</td>
</tr>
<tr>
<td>( D_{\text{series}}(K_B=K_4) )</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>( K_{B+1} - K_B )</td>
<td>( K_{B} - K_B =0 )</td>
</tr>
<tr>
<td>...</td>
<td>...</td>
<td>...</td>
<td>...</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>( D_{\text{series}}(K_B=K_k) )</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

This table shows a generalized example of how we designated each image as the baseline in turn, and calculated the amount of decorrelated pixels across the whole data set.
List of Terms

$l = \text{a time step in the decorrelation analysis} \ (l = 1, 2, 3, \ldots, k)$

$k = \text{The total number of time steps involved in the decorrelation analysis}$

$K_l = \text{The } l^{th} \text{ binary wet-dry matrix in the time series from the experimental images.}$

$t = \text{experimental runtime}$

$t_l = \text{The experimental runtime of the } l^{th} \text{ time step in the time series wet-dry matrices}$

$D = \text{The bulk decorrelation, or total number of pixels that have changed between two}$

binary wet-dry matrices.

$D_{\text{inst}} = \text{The instantaneous bulk decorrelation between two binary wet-dry matrices;}$

“instantaneous” means that the two matrices are immediately successive, and can therefore be denoted $K_l$ and $K_{l+1}$.

$\Delta t_{\text{inst}} = t_{l+1} - t_l = \text{the amount of time elapsed between two subsequent time steps; this is}$

used to scale instantaneous decorrelation such that it is in units of decorrelation/time.

$i = \text{the } i^{th} \text{ row the binary wet-dry matrices}$

$j = \text{the } j^{th} \text{ column in the binary wet-dry matrices}$

$m = \text{The number of rows in the wet-dry matrices}$

$n = \text{The number of columns in the wet-dry matrices}$

$K_B = \text{The baseline wet-dry matrix, to which the whole time series of transient matrices}$

$(K_T)$ are compared

$K_T = \text{The transient wet-dry matrix, where } T = [B, B+1, B+2, \ldots, k]; \text{ it is compared to}$

matrix $K_B$ to find the level of correlation between the itself and $K_B$

$D_{\text{series}} = \text{Bulk decorrelation along a time series; this is the total number of pixels that have}$

changed between matrices $K_B$ and $K_T$
Chapter 3: First-Order Observations and Advanced Methods

3.1. Introduction

In Chapter 2, we develop a method with which to measure bulk pixel decorrelation in a series of images over time. The next step is to apply this method to understanding how fluvial systems behave. In particular, we use bulk decorrelation with respect to a baseline time step, B, to develop a channel time scale. This shows us how rapidly the similarity between a baseline time step and a subsequent time step decays over time, and allows us to define a point at which the new channel pattern has no obvious relation to the channel pattern in the baseline time step. This is enough information to construct a channel-mobility time scale that we can apply to fluvial systems.

In order to place bulk decorrelation in a better context for channel networks, we develop two separate scaling parameters. Then, we subtract the new, scaled decorrelation from 1. This gives the general form given in Equation 3.1 (where \( D_{\text{series}} \) is the decorrelation in a series of images compared to some baseline image and SP is the scaling parameter)

\[
C = 1 - \left( \frac{D_{\text{series}}}{\text{SP}} \right)
\]  

(3.1)

Since 1 minus the scaled decorrelation gives the opposite of decorrelation, this resulting parameter is called correlation, represented by the letter C. Here, “correlation” refers to “correlation” in its standard English-language sense. It is not rigorously
correlation in a statistical sense; hence we call the resulting value “correlation” instead of the “correlation coefficient”. However, the general idea of “how many pixels are the same” applies correlation-like ideas to river channels instead of functions.

3.2. Constant Scale: Flume Surface Area

This first scaling parameter we applied was a constant scale: the size of the image. This is simply defined as $A = m \times n$, where $A$ is the total image area in pixels (i.e., the total number of entries in the matrix), $m$ is the number of rows in the matrix, and $n$ is the number of columns. In order to obtain correlation scaled to image area, the following equation was applied

$$C_A = 1 - \left( \frac{D_{\text{series}}}{A} \right)$$  \hspace{1cm} (3.2)

where $C_A$ is the area-based correlation, which is just one minus the area-scaled decorrelation.

The advantage of using this method is that the size of the image is the size of the subaerial flume surface area; therefore, the result is the number of pixels changed divided by the total area of possible pixel change. In addition, this scaling parameter is constant. Therefore, it provides a datum against which further analysis (see section 3.3) with variable scaling parameters can be measured.
3.3. Random Scale

The second scaling parameter applied to the decorrelation outputs a correlation plot such that a correlation value of 1 represents perfect correlation, such that every pixel is in the same location in both images, and a correlation value of 0 means that the amount of correlation between these two images is equal to the correlation of a random scatter of wet and dry pixels, in the same proportions as are given by the images being compared. This scaling is more insightful because (1) random scatter is a good baseline against which to compare a system that is decorrelating from its original position and (2) looking at values greater than or less than zero is a natural way to analyze a system, and adding an intrinsic meaning to zero helps this. We use this scaling to provide almost all of the results presented in this thesis.

3.3.1. Reasons to Scale to Randomness

One major goal of this study is to quantify how quickly a channel system decorrelates to the point at which it has no memory of its initial form. Therefore, we need a quantitative way to decide when the point is reached at which the initial and current channel positions are completely unrelated. We apply a value of randomness as the value for zero correlation, as we postulate that a system that is correlated with itself in the same amount as a random scatter is not correlated with itself at all. This is illustrated below in the argument for “accidental recorrelation”.

If flume subaerial surface area is the only scaling parameter, there is no way for the correlation to reach zero unless the wet and dry pixels are in a perfect 50-50 balance. In
most of the experiments, wetted area is between 20 and 35 percent. Therefore, the maximum number of pixels that could decorrelate is 40%–70% of the total number, respectively. In these cases, area-scaled correlation would never go below 0.6 or 0.3, respectively. This is because the maximum number of pixels that can possibly decorrelate in a binary system is 2 times the number of pixels of the type (i.e., 0 or 1) that is in the lower quantity. For example, in a 3-by-3 matrix in which we have four zeros and five ones, only four zeros can switch places with the ones, leaving the last “one” in its original place.

Since we prefer a completely decorrelated system to have zero correlation, and because the amount of decorrelation would then also be related to the number of wet pixels in the system, this scaling to subaerial surface area is not the most desirable of scaling parameters. Therefore, an attempt was made to scale the images by the total number of pixels that could change. However, this scaling was unsuccessful because it created a bias in the data and because it is not natural for the fluvial system to abandon all of its original wet pixels (next paragraph). Therefore, we applied the random scaling with more success.

One of the advantages of random scaling has to do with “accidental recorrelation”. In a flume of limited area (or in a real delta or riparian corridor, for that matter), the probability that the whole channel system would have every one of its originally wet pixels become dry is vanishingly small. In reality, if the plan form of a channel system has nothing to do with the plan form at a different time step, there is still some correlation because of reaches of channels that “accidentally” cross one another (Figure 3.1).
Therefore, scaling to randomness provides a method by which decorrelation analysis, which is blind to specific channel morphology and directionality, can reach zero at the point at which “accidental” recorrelations occur at a density that is representative of randomness. In our theory, this point is representative of the point at which the two plan forms are no longer related.

**Figure 3.1.** “Accidental recorrelation”. Images A and B are images from the braided and vegetated dynamics experiment (Tal and Paola, 2007) selected far enough apart that there is little relationship in their plan forms. Image C is a superposition of images A and B to illustrate accidental recorrelation, regardless of original plan forms. In images A and B, white is wet and black is dry. In image C, black is dry in both images, white is wet in both images, and gray is wet in one image and dry in the other. Despite the differences between images A and B, the areal extent of similarity outweighs the areal extent of change in image C.
3.3.2. Construction of the Random Scaling

In order to scale to randomness, we divide the bulk decorrelation by the amount of decorrelation that would be experienced by a given channel network becoming a random scattering of wet and dry pixels. Using the probability that a given pixel would change, based on the fractions of wet or dry pixels in the baseline and transient matrices, the value for randomness was defined as

$$\Phi_D = f_{w,B} \cdot f_{d,T} + f_{d,B} \cdot f_{w,T}$$

(3.3)

where $\Phi_D$ is the average decorrelated fraction among randomly scattered pixels, $f_{w,B}$ is the fraction of pixels in the baseline image that are wet, $f_{d,T}$ is the fraction of pixels in the transient image that are dry, $f_{d,B}$ is the fraction of pixels in the baseline image that are dry, and $f_{w,T}$ is the fraction of pixels in the transient image that are wet. When correlation was scaled to this value of randomness, the resulting equation became

$$C_{\Phi} = 1 - \left( \frac{D_{\text{series}}}{A} \right) \left( \frac{\Phi_D}{\Phi_D} \right)$$

(3.4)

where $C_{\Phi}$ is the randomness-based correlation, $D_{\text{series}}/A$ is the area-based decorrelation, and $\Phi_D$ is the average decorrelated fraction as defined above.

The key to this idea is that when a river no longer has any memory of its original plan form, the amount of scaled correlation hovers around zero, or a higher constant for
systems with increased forced coherence to their former shapes. Fluctuations above and below that value show specific changes in channel plan form with respect to the location of the channel in the baseline time step. In addition, this randomness-based correlation is not significantly different than the area-based (or flume-surface-based) decorrelation in shape: it just arrives much closer to zero (see Figure 3.2). Therefore, the randomness-scaled decorrelation was deemed to be a safe method to derive additional insight into the change in river channel patterns with time.

A further advantage of this method is that, since it is sensitive to the amount of wet and dry pixels, it balances out changes in the channel system in a much more naturally interpretable way. For example, if a system has only one small channel which gives a wetted area of 10%, the scaling value (ΦD) entered into Equation 3.5 is much smaller than that for a system with a 50-50 balance of wet and dry pixels. Therefore, two events that change the same number of pixels can cause a different change in correlation within the system. For example, in two events that change the same number of pixels, what would be interpreted as large event for the system, such as the complete avulsion of the small channel, will make a much larger impact on correlation the system than what would be interpreted as a smaller event, such as migration or flow expansion in the 50-50 case. Because of these advantages, randomness is the primary method for correlation scaling that is performed throughout the rest of this thesis.
Correlation with Time
DB03-2: Runtime Hours 833-1580 (baseline image at 833 hours)
Scaled to Total Pixels (red, dashed) and Randomness (blue, solid)

Correlation Scaled to Total Number of Pixels (Floodplain Area)
Correlation Scaled to Randomness (Random Scatter of Pixels)
List of Terms

$l$ = Experimental time step
$B$ = Baseline time step
$T$ = Transient time step
$K_l$ = The $l^{th}$ binary wet-dry matrix in the time series of experimental images.
$K_B$ = The baseline binary wet-dry matrix, to which the whole time series of transient matrices ($K_T$) are compared
$K_T$ = The transient wet-dry matrix, where $T = [B, B+1, B+2, \ldots, k]$; it is compared to matrix $K_B$ to find the level of correlation between the itself and $K_B$
$C$ = The fraction of correlated pixels, or pixels that are the same between time steps, when the bulk decorrelation along a time series ($D_{series}$) is scaled by some scaling parameter.
$D_{series}$ = Bulk decorrelation along a time series; this is the total number of pixels that have changed between matrices $K_B$ and $K_T$
$SP$ = The scaling parameter by which the bulk decorrelation along a time series ($D_{series}$) is normalized
$A = m \times n$ = The number of entries in a binary wet-dry matrix with dimensions $m$ by $n$; this is also the area (in pixels) of the original image that produced the binary wet-dry matrix.
$m$ = The number of rows in the wet-dry matrices
$n$ = The number of columns in the wet-dry matrices
$C_A$ = The fraction of correlated pixels, scaled to total area, "A"
$\Phi_D$ = The average decorrelated fraction among randomly scattered pixels in two binary wet-dry matrices, where $f_{w,B}, f_{w,T}, f_{d,B},$ and $f_{d,T}$ are the respective fractions of wet and dry pixels in the two binary wet-dry matrices being considered.
$f_{w,B}$ = The wet fraction (fraction of pixels that are wetted) in matrix $K_B$
$f_{w,T}$ = The wet fraction in matrix $K_T$
$f_{d,B}$ = The dry fraction in matrix $K_B$
$f_{d,T}$ = The dry fraction in matrix $K_T$
$C_{\Phi}$ = The randomness-scaled correlation, or fraction of correlated pixels, scaled to $\Phi_d$
Chapter 4: The Theory behind the Analysis

4.1. Introduction

In order to apply the decorrelation time scales most effectively, this section develops general theories behind the analysis developed in Chapter 3. Here in Chapter 4, we first describe the exponential decay shape of the curves of correlation with time (4.2), their various properties (4.3 and 4.3.1), and the meanings of the terms in the exponential decay equation (4.3.2, 4.3.3, and 4.3.4). Then, a series of thought experiments and numerical experiments which bear more immediate relevance to river channels are investigated (4.4). These illustrate the significance of the exponential decay curves on the kinematics of the channel systems. After this, we investigate constant-in-time variations in the curves of randomness-scaled correlation with time (4.5) because these contain signals of particular positions occupied by the river channel at a certain time. These signals of channel position are especially strong in the experimental deltas, where one can find clear indicators of the time scales in which channels occupy certain positions on the delta surface. After this, we use spatial decorrelation to analyze a system of Lagrangian coherent structures in turbulence in an attempt to investigate the generality of exponential decay with decorrelation (4.6). These Lagrangian coherent structures are the constantly-deforming attractors and repulsors which determine the velocity field within a turbulent fluid, and our plots show that the migration of these Lagrangian coherent structures are analogous to the river channels in that they also experience exponential decay with time, with respect to a baseline time step. This suggests the more general existence of exponential decay in spatial correlation with time, within natural systems. Finally, we
quickly state our method for comparing channel migration rates by comparing the decay constants of the decorrelation profile for different fluvial systems (4.7).

### 4.2. First-Order Features of “Correlation with Time” Curves

The first two significant features of a series of plots of correlation with time (under any scale) are that (1) correlation falls off with time as an exponential decay function and (2) that the rate and magnitude of the decrease in correlation with time remains approximately steady regardless of which time step is chosen as the baseline, so long as the experimental boundary conditions remain unchanged (Figures 4.1 and 4.2). In addition, individual discrepancies in the data that push it away from a perfect exponential decay remain constant in time. This leads to the following important fact: The general shape of the exponential decay curve remains constant in time relative to each baseline time step (Figure 4.1). Therefore, the exponential decay can be thought of as *constant in relative time*. Individual autogenic fluctuations are events marked in time, and therefore remain placed at a given time step without general regard to the baseline image that is being chosen (Figure 4.7). These events are *constant in absolute time*, and are markers of change in overall channel form. In other words, a series of decorrelation profiles plotted with their baseline time step at $t = 0$ approximate a characteristic exponential decay curve for the system, which gives a general time-averaged channel mobility. On the other hand, events that are constant in absolute time show how the fluvially active area moves in a way that is specific to the particular experimental runtime.
Figure 4.1. **Exponential decay of correlation profiles.** (Experimental data are from Martin et al., manuscript in preparation, 2007.) This figure from three randomly selected time-steps shows that plan form (wet vs. dry area) correlation with time for experimental fluvial systems, with respect to any baseline time step, roughly follow the shape of an exponential decay curve.
4.3. The Exponential Decay Function, and its Properties as Applied to the Plots of Correlation with Time

4.3.1 Exponential Decay: Constant in Relative Time

Correlation with respect to time from the baseline time step is described by the exponential decay function (Figures 4.1 and 4.2 and Section 4.2), which is given by the equation (with the symbols for constants here labeled as used in the remainder of this thesis):

\[ C_\phi = ae^{-M \tau} + c \]  

(4.1)

where \( C_\phi \) is the randomness-based correlation and is the dependent variable. Constants \( a \), \( M \), and \( c \) are empirical, and describe the equation particular to the specific system being analyzed. The term \( \tau \) is the independent variable: time. For these experiments, the data are arranged such that \( \tau = 0 \) when the baseline image \( (K_B) \) is itself being analyzed (and the randomness-scaled correlation, \( C_\phi \), is therefore equal to 1). This is due to the fact that, with the same boundary conditions, the curves relative to each baseline time-step are self-similar.

4.3.2. Meaning of the Constants, Part 1: The “Confinement c”

The constant \( c \) gives the proportion of correlation at the horizontal asymptote. This is called the “asymptote of randomness”, because it stays constant except for small wiggles that are created when the transient channel plan form crosses the initial (baseline) plan.
form in a greater or lesser way, without being morphologically similar. This value is zero for a completely unconfined system, where the fluid is free to decorrelate to a state in which it has no memory of its initial position. (A crude example for a behavior like this could be a random walk, although it is a simplification in that channel morphology is deterministic with respect to fluid dynamical, depositional, and topographic conditions.) When a channel system is not able to move across the entire area under analysis, \( c \) has a value greater than zero. This occurs in systems with some form of fluvial confinement that is either permanent or evolves on a time scale much longer than a channel-migration time scale. Examples of these features are incised valleys from allogenic forcings (as seen in the experimental work of Kim et al., 2006; Kim, 2007), areas that are created by more drastic events than the currently operating fluvial processes (as seen in our analysis of the work of Tal and Paola, 2007), and analyses that view a part of the deltaic surface too close to the water and sediment feed (as seen in our analysis of the low-Froude-number experiment of Martin et al., 2007). The specific value of \( c \) is determined by the amount by which the flow is confined.

4.3.3. Meaning of the Constants, Part 2: “\( a \)”

It could be expected that, because correlation starts at \( t = 0 \) and \( C_\Phi = 1 \), \( a \) should be equal to \( 1 - c \). However, this is not the case. We hypothesize that the problem is that the use of an arbitrary threshold between “wet” and “dry” causes a certain number of pixels that are on the border of being wet or dry to move from one classification to the other from time step to time step because of small changes in lighting, dye concentrations, or other parameters. Therefore, the degree of decorrelation between the first and second time steps
is unexpectedly large. After the second time step, this extra decorrelation has already
been accounted for, so decorrelation occurs here at a rate that decreases at a consistently
with time (the definition of exponential decay). Because of these reasons, \( a \) has a value
such that \((c+a) < 1\).

4.3.4. Meaning of the Constants, Part 3: "M"
The decay constant \( M \) is the most important to our analysis, in that it gives the rate of
decorrelation (in units of 1/time). We therefore use \( M \) as our characteristic channel
mobility. In order to compare the channel mobilities of two different experimental data
sets, we compare the values for \( M \) in their characteristic decorrelation profiles. (However,
we first normalize the values of \( a \) and \( c \) in these profiles, as a precaution against seeing
the effects of \( a \) and \( c \) on the values for \( M \).)
Figure 4.2. Correlation versus time in an experimental delta. This figure shows correlation with time for runtime hours 1800-2400 of the Low-Froude-Number Experiment (DB03-2). (Experimental data are from Martin et al., Manuscript in Preparation, 2007.) Time is measured from each baseline time step, taking advantage of the self-similarity in the decorrelation from any initial time step (so long as the boundary conditions remain the same). In this way, a scatter plot is generated and the trend is measured.

This figure shows the importance of the “confinement c”. The delta, unbounded by immobile obstructions, decorrelates to near randomness (0). (It does not completely decorrelate to randomness because the water and sediment feed is locked in place, and this imposes a confinement on where the flow can be when it is near the water and sediment feed.)
Correlation vs. Time
2002 Run of the Braided and Vegetated Dynamics Experiment
Nonvegetated and Fully Braided Segment

$C = 0.6037\exp(-1.670t) + 0.2151$

Figure 4.3. Correlation versus time for the braided, non-vegetated portion of the Braided and Vegetated Dynamics Experiment. (Experimental data are from Tal and Paola, 2007.) It is formatted in the same manner as figure 5.1, above. For this curve, $R^2 = 0.9067$.

The "confinement c" is approximately 0.21, well above randomness. This is hypothesized to be because of flow deflection. The side walls in the flume can be compared to a bedrock valley in which the braided stream is located: these "side walls" in this field example are modified on a time scale that is much longer than a channel time scale. In addition, the bedrock valley could be created by more powerful events than the stream could provide; therefore, the dimensions of the valley may not be optimal to allow the stream to flow across its whole surface area with approximately equal probability. Deflections in flow direction because of these features (both in the experiment and in the field example) would cause the channel system to rework only part of its area, and would therefore raise the "confinement c".

The gap between time-steps 2 and 3 is due to missing data.
4.4. Understanding Exponential Decay in Terms of River Channels

4.4.1. An Oversimplified Thought Experiment about Channel Migration

In a simple thought experiment about channel migration, a channel can be represented by a rectangular body of water, moving in one direction with a certain velocity. The decorrelation of the rectangle over time is linear, and it drops to zero at the instant the migrated rectangle no longer has any overlap with the original rectangle.

4.4.2. A Simple Numerical Model of a Migrating Channel Bend

A second simple experiment is to compare a scenario that is more natural to a river system, such as a bend in a channel. In a simplified sense, bends tend to migrate outward. Therefore, we modeled a series of parabolas of finite width and gradually increasing sinuosity over time. During each time step, the outer bend of the sinuous element (i.e., the apex of the parabola) migrated a fixed distance outward. The reason for keeping this rate constant is that many field measurements yield a constant rate for the migration of the outer bend of an element of sinuosity.

Figure 4.4 shows a number of different lines, each of which represents a different baseline time step. When the original image has more curvature, rate of decorrelation slows and the asymptote is higher. This reflects the fact that comparing stretched parabolas is nonlinear, and that the angle relationships within a stretched parabola show that the difference between two time steps decreases with a greater degree of initial stretching.
Our main findings were that (1) the decorrelation was roughly linear, and flattened off sharply when the bend had migrated one channel width, and (2) the greater the curvature of the bend at the baseline time step, the shallower the slope of the decorrelation (fig. 4.4). Both of these results are as expected for the curved geometry. Although this is a simple model, it does show additional evidence for the relationship between nonlinear decorrelation and channel migration.

4.4.3. A More Complex Thought Experiment: A Series of Migrating and Avulsing Rectangles

In a more complex thought experiment, the experimental system exists of rectangles that both migrate and avulse. These rectangles, on average, will migrate linearly away from their starting position, and at some time, will avulse and be in a different part of the system until they either migrate back or avulse back to their original position. In general, these plots will have a linear component with negative slope, followed by a flat line at zero with a few wiggles. Averaging a large number of these graphs with the proper avulsion frequency would seem to give the correct profile (Chris Paola, personal communication). However, the fact that the exponential decay is seen in non-avulsing environments (e.g., the work of Tal and Paola, 2007) shows that, although this is another way to try to understand the system, it is also incomplete. In the end, it is probably a combination of migration, avulsion, channel morphology, and the stochastic nature of river-channel motion that results in the fact that channel migration can be approximated as an exponential decay (Section 3.3.1, Figure 3.3, Section 4.4.4).
Figure 4.4. Channel-bend simulation: parabola stretching. (above) Randomness-scaled correlation vs. time. The correlation does not reach zero because it was measured using a study area much larger than the migration distance of the channel system. This is analogous to a channel which is prevented from touching certain areas on its floodplain (Figure 4.2).

Figure 4.5. A Visual Representation of Some Time-Steps from the “channel-bend” Parabola Model. This figure shows the parabola’s position at several time-steps. The darkness of the image shows density of pixel occupation over time, with black being the most occupied.
4.4.4. Definitions of the Regimes within the Exponential Decay

The exponential-decay curve can be thought of in three zones: linear migration, pseudorandomness, and a middle transition zone. For time steps close to \( l = B \), the channel positions are extremely close to those in the baseline image. Therefore, channel migration is approximately linear: all channel motion acts to decorrelate an approximately even number of pixels from the original plan form. This can be thought of in terms of the moving rectangle (Section 4.6.1) or in terms of the stretching bend in the channel (Section 4.6.2) before the inner edge of the bend in the transient time step passes the outer curve of the bend in the baseline time step. However, the exponential decay tells us that this rate is constantly decreasing, so the farther a channel is from its location in the baseline time step, the less rapidly it is moving away from its original position. When the transient time step is at a time far from the baseline time step, the plan forms of the channels have no obvious relationship to one another. In this case, the time-averaged relationship to the baseline time step is unchanging with continuing time away from the baseline. This is called the asymptote of randomness, in that all correlation is caused by the random crossing of the plan forms at the baseline and transient time-steps (see “accidental recorrelation”, Figure 3.3). The time period in between these two approximately linear states is called the transition zone. During the time steps constituting the transition zone, the channel form is not completely unrelated to the channel form in the baseline time step, but it is not very similar either. Therefore, both the roughly linear migration and the asymptotic state of randomness are “acting” on the decorrelation, and it is part-way between both states. These states are noted in the Figure 4.6.
Figure 4.6. A river-based way of looking at correlation. This plot highlights zones of linear migration and decorrelation, transition, and randomness. It is of correlation (randomness-scaled) vs. time in the low-Froude-number-experiment (hours 1800 to 2400).

This figure illustrates the three main zones within a typical plot of randomness-scaled correlation with time: the zone of quasi-linear migration and decorrelation, the transition zone, and the asymptote of randomness.
4.5. Variations Showing Channel Positions: Constant in Absolute Time

The curves of correlation with time are not smooth. In the deltaic experiments in particular, these plots are filled with abrupt jumps up and down in correlation with time, and these jumps stay in the same location without regard to the baseline time step being chosen. These features of the data on correlation with time that deviate from an exponential decay curve are static in absolute time: they occur at the same place regardless of which baseline image is chosen for analysis. This is as opposed to the exponential decay curve, which is constant in relative time: with respect to the time steps of their respective baseline images, curves of correlation with time follow approximately the same exponential decay shape. (This is illustrated in figures 4.1, 4.2, and 4.3.)

These features show major shifts in the position of the channels. In a deltaic setting, such as DB03-1 (which we examine here), migration and/or avulsion of main-branch channels on deltas to a previously unoccupied region of the delta surface can cause these (Sheets et al., in press, 2007). Here, each jump represents separates regimes of main channel position. In a more terrestrial setting, these features show avulsions in a mature single-thread, sinuous channel system (Tal and Paola, 2007).

An important feature of constant-in:absolute-time variations in the plots of correlation with time is their potential use to understand channel-reoccupation time scales, and what this means in terms of delta evolution (Perignon, Wickert, and others, manuscript in preparation, 2007). Analysis of the data for correlation with time (Figure 4.7) and the corresponding positions of wet and dry surface area (Figure 4.8) show that the left and
right sides of the delta in DB03-1 serve as the focus of the flow during each of these regimes. Experimental data from Sheets et al. (in press, 2007) show clusters of channel deposits in the stratigraphy and show that the channels themselves periodically switch between occupying the left and right sides of the delta. This autogenic cyclicity is remarkable, and worthy of study in terms of the its geomorphic, sedimentologic, and stratigraphic causes and implications.

We do not further address this cyclicity here, but we anticipate the work of Sheets et al. (2007) to have a more complete discussion of autocyclic channel clustering. In future work (e.g., Perignon, Wickert, and others, manuscript in preparation, 2007), we plan to investigate features of the topographic surface that allow certain areas of the deltaic surface to repeatedly serve as attractors for the flow, such as those that are shown to exist in the work of Sheets et al. (2002).
Figure 4.7. Illustration of constant-in-absolute-time features in DB03-1; 01/23 run; timesteps 1-60. This figure was created by plotting randomness-scaled correlation against time for several baseline time-steps, and placing each of these baseline time-steps at its experimental runtime (instead of at $t = 0$). Therefore, the values on the $x$-axis are absolute experimental runtimes, instead of time referenced to the runtime of the baseline time-step. Jumps in the plot, such as the one at approximately 11 hours, 10 minutes, exist without regard to which baseline time step has been selected for the analysis, and signify motion in the locus of fluvial activity on the surface of the delta.

The horizontal line signifies the expected asymptote of randomness. The value for correlation is above the asymptote of randomness when the flow is on the same side of the delta as in the baseline time step, and is below randomness when the flow is on the opposite side of the delta than in the baseline time step.

The numbers across the top signify the position of the image numbers from Figure 4.8 on the following page on this graph.
**Figure 4.8** (Previous Page) Channel positions with time, DB03-1. This plot shows pictures of channel positions with time, referenced to positions on the graph in Figure 4.7. This figure is intended as a visual tool for understanding the relationship between channel position and constant-in-time features on graphs of correlation with time. Each time step is given a number (which references the descriptions given here and its the location on the graph in Figure 4.7) and a time (in hh:mm:ss). Note that, because the flow was on the left in the initial conditions, correlation at time steps far from the initial time step rises above randomness when flow is on the left and sinks below randomness when flow is on the right.

The following are descriptions of the channels on the deltas with the corresponding numbers and at the times given above. In these descriptions, “left” and “right” are taken to be the viewer’s left and right (not river-left and river-right).

1: The channels flow on the left side of the delta.
2: A short-lived attempt to occupy the right-hand-side of the delta with flow fails. You can faintly see shallow flow in the channels on the right-hand side of the delta.
3: Flow continues to occupy the left side of the delta.
4: Flow is now occupying the full delta surface.
5: The flow has moved to the right side of the delta. You can see the remains of a small channel on the left that is just about finished filling its former path with sediment.
6: Thirty-five seconds after image 5, flow has abandoned the left side of the delta completely.
7: Flow continues to occupy the right side of the delta.
8: The flow is still on the right, though the channels are moving centerward.
9: The channels are once again occupying the entire delta surface.
10: Flow now reoccupies the left side of the delta.
4.6 Additional Evidence for Exponential Decay of Correlation in Nature: Lagrangian Coherent Structures in Turbulence

4.6.1. Introduction

Exponential decay is found in many places in nature, including radioactive decay (Rutherford, 1902), thermal conduction (Fourrier, 1822), gas diffusion (Einstein, 1906), the density of the atmosphere with distance from the earth, and any system that can be simulated by a random walk. Therefore, we are interested in looking deeper into the generalities of the exponential decay that we see in the correlation with time of the river channels. Paola (1996) has suggested a link between turbulence and channel pattern. In particular, he established a relationship between a braided stream and fluid turbulence by successfully describing braided morphologies through turbulence equations. Because of this relationship to river channels and because we know that turbulence is a stochastic phenomenon, we would like to investigate whether looking at the movement of attractors and repulsors in turbulence over time also behaves as an exponential decay function. The particular system of our investigation is the motion of the Lagrangian skeleton of turbulent flow, recently visualized by Mathur et al. (2007).

Here, a rough image set of new experiments showing the Lagrangian coherent structures in turbulence (Mathur et al., 2007) is analyzed for bulk pixel decorrelation. These Lagrangian coherent structures are defined lines or surfaces within the Lagrangian—particle-based, as opposed to the Eulerian (observer-based)—frame which either attract or repel particles (Haller and Yuan, 2000). These Lagrangian coherent structures serve as
the locus for alteration within a fluid flow, and are therefore useful for obtaining a deeper knowledge of geophysical processes (Haller and Yuan, 2000). The attractors are local zones of stability within the turbulence, and the repulsors are local zones of instability (Haller and Yuan, 2000). In the rough, image-based data set that was analyzed, it was not possible to separate the attractors and repulsors. However, the bulk motion of attractors and repulsors throughout the flow as a whole can still serve to show a first-order picture of how the foci within the turbulent structure change in location with time. The resulting analysis shows that the correlation with time for the Lagrangian coherent structures as a whole can be represented by an exponential decay curve. This is like the fluvial experiments, except in that the exponential decay curve for the randomness-based correlation Lagrangian coherent structures is much closer to a simpler exponential decay, \[ C_\phi = e^{-Mt} \] (4.2) than the fluvial systems, which require the additional coefficients \( a \) and \( c \): \[ C_\phi = ae^{-Mt} + c \] (4.3)

4.6.2. Experimental Methods

The following methods are paraphrased from those given in the work of Mathur et al. (2007). The experiments were conducted in a tank 50 cm high and 40 cm in diameter, rotating at 0.4 Hz. Water was pumped into and out of the bottom of the tank, which
produced turbulent flows with Reynolds number (Re) 6000 and Rossby Number (Ro) 20. In the higher reaches of the tank, the flow shows quasi-2D turbulence, and it was this near-surface turbulence that was analyzed for Lagrangian coherent structures. In this surface flow, Re is 1000 and Ro is 0.3 (Mathur et al., 2007).

4.6.3. Results

The resulting plot of randomness-based correlation with time shows the general form of the data, along with an associated curve. There is some concern about the step-function nature of this result, but it is believed that the data points at time $t = 0.2$ seconds displays good evidence of the kind of curvature necessary for exponential decay. It is believed that, were the sampled data set selected at shorter intervals, time steps such that $0 < t < 0.2$ seconds would show additional evidence for an exponentially decaying profile.

One major feature of the decorrelation of Lagrangian coherent structures in this experimental system, as opposed to experimental fluvial systems, is that the Lagrangian coherent structures decorrelate on the scale of hundredths to tenths of a second, while the experimental fluvial systems decorrelate on the scale of minutes to hours. In addition to scaling differences between these two types of experiments, this discrepancy in decorrelation time is likely because sediment imposes a critical boundary shear stress requirement for plan-form change, which makes decorrelation much harder and takes a longer time than simple turbulent mixing.
Correlation with Time
Lagrangian Coherent Structures in Turbulence
(Considering Attractors and Repulsors Together)

* Lagrangian Coherent Flows
- $C = 0.9921 \exp(-1.656t) + 0.007938$

$u_0 = 5, 10, 15, 20$
$0 = 5, 10, 15, 20$

Time (tenths of a second)

Therefore, more data between time = 0.0 seconds and time = 0.2 seconds would be needed to confirm this hypothesis.

Figure 4.9. (Left) (Figure from Mathur et al., 2007.) Visualization of Lagrangian coherent structures. This figure shows the Lagrangian coherent structures, or structures within turbulence that drive the turbulent flow. The red lines represent curves that attract fluid particles, and the blue lines represent curves that repel fluid particles. Both of these curves evolve through time, and we believe that the movement of these curves together can be described by an exponential decay function (figures 4.9 and 4.10). Image b is a blown-up version of the area inside the heavy-bordered black square.

Figure 4.10. (Below) Randomness-scaled correlation vs. time for the bulk movement of repulsive and attractive Lagrangian coherent structures. (By “bulk movement”, we mean that the attractors and repulsors were not differentiated.) This plot produces an exponential decay curve similar to those for the experimental river channels. However, this “exponential decay” result is highly dependent on the data at time = 0.2 seconds.
Figure 4.11. Randomness-scaled correlation with time for the Lagrangian coherent structures, plotted on a logarithmic y axis and a linear x axis. Note that the asymptote is only a small value above zero. This is hypothesized to be due to minute errors in our analysis, and that it should equal 0 for full diffusive (turbulent) mixing.
4.7. Developing a Metric for Correlation-Based Channel Time Scales

4.7.1. Decay Constants

Differences in channel migration rate are measured by comparing two similar fluvial systems that have a small number of distinct extrinsic parameters. The rate of channel mobility in these plots is best represented by \( M \), the decay constant. This decay constant has units of (1/time). Because the \( a \) and \( c \) values are usually close to one another in the experiments that are compared, these values generally are set to remain constant at the average of these values for all of the data being analyzed, thus leaving \( M \) as the only term that changes.

4.7.2. The e-Folding Time Scale

In this study, a characteristic channel time scale is given by the the e-folding time, or time at which the correlation reaches a value of \( (a/e) + c \). (The terms \( a \) and \( c \) are as given for the exponential decay, and \( e \) is the base of natural logarithms.) The values for \( a \) and \( c \) can be either those for the specific curve, or the slightly altered values used for comparison with other analyses.
List of Terms

t = time, usually relative to the baseline time step (t = 0 when K_f = K_B)

C_φ = The randomness-scaled correlation (when referring to the exponential decay function instead of the data, this value is taken to be the temporally averaged randomness-scaled correlation)

a = The constant in the exponential decay function that determines how much the function is vertically stretched.

M = The decay constant in the exponential decay function, which determines relative decorrelation rates as well as the curvature of the plot.

c = The constant in the exponential decay function that determines the location of the horizontal asymptote. When used with random-scaled data, c determines the level of permanent confinement on the flow in the system, which is the amount of flume area that the channel does not cross.

Re = The Reynolds number, a ratio of inertial forces to viscous forces in the fluid

Ro = The Rossby number, also known as the Kibel number, is a ratio of fluid acceleration over Coriolis acceleration
Chapter 5: Descriptions of the Experiments
(Parameters and Boundary Conditions)

5.1. Introduction

Time-lapse images of fluvial-system experiments were analyzed for channel migration by separating channel (colored with dye) from their substrate. Four of these experiments were deltas (Martin et al., manuscript in preparation, 2007; Sheets et al., in press, 2007; Hoyal and Sheets, manuscript in preparation, 2007; Kim et al., 2006; Kim, 2007). The final experiment under analysis was a braided river that was eventually forced into a single thread by simulated vegetation (alfalfa: *Medicago sativa*) (Tal and Paola, 2007).

All of these experiments were performed at the University of Minnesota’s Saint Anthony Falls Laboratory. Relevant experimental parameters are listed in the text. For ease of readability, these experiments may be referred to by their boldface nickname or nicknames when appearing in the text of the remainder of this paper, without the corresponding citation that is noted here. Therefore, we would like to assign due credit now for the work that has been done. The experiments were:

- The 2002 Run of the Experimental EarthScape “Jurassic Tank” Facility (*XES02*) (Kim et al., 2006; Kim, 2007)
- The 2002 Run of the Riparian Vegetation and Braided Stream Dynamics Experiment (*Riparian Vegetation and Braided Stream Dynamics* or *Braid/Veg*) (Tal and Paola 2007)
- The First 2003 Run of the Delta Basin (DB03-1): high-resolution topography experiment (High-Res, DB03-1, or HR) (Sheets et al., in press, 2007)
- The Second 2003 Run of the Delta Basin (DB03-2): low-Froude-number experiment (Low-Fr, DB03-2, or LF) (Martin et al., manuscript in preparation, 2007)
- The 2007 cohesive sediment experiment (Cohesive Experiment or 2007 cohesive experiment) (Hoyal and Sheets, manuscript in preparation, 2007)

The sedimentary substrates across all of the experiments were composed of sands of different types, clays, a special mixture of cohesive sediment (Hoyal and Sheets, manuscript in preparation, 2007), and alfalfa (Medicago sativa) as simulated vegetation.

The specifics of the sediment types used are illustrated in the following table:

**Table 5.1. Sediment types used for each of the analog experiments**

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Sediment Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>DB03-1</td>
<td>70% quartz sand (0.120 mm); 30% bimodal anthracite (0.190 and 0.460 mm)</td>
</tr>
<tr>
<td>DB03-2</td>
<td>70% quartz sand (0.12 mm); 30% bimodal anthracite (70% fine (0.10 mm) and 30% coarse (0.50 mm))</td>
</tr>
<tr>
<td>Cohesive Experiment</td>
<td>An approximately uniform silicilastic mixture that ranges from clay to coarse sand, with a small amount of commercially available polymer to add cohesion</td>
</tr>
<tr>
<td>XES02</td>
<td>63% quartz sand (0.110 mm), 27% bimodal anthracite (0.460 and 0.190 mm), and 10% kaolinite</td>
</tr>
<tr>
<td>Braid/Veg (nonvegetated stage)</td>
<td>Pure quartz sand, D₅₀ 0.5 mm</td>
</tr>
<tr>
<td>Braid/Veg (vegetated stage)</td>
<td>Pure quartz sand, D₅₀ 0.5 mm, in which alfalfa (Medicago sativa) is growing</td>
</tr>
</tbody>
</table>
All of the sediment types moved primarily as bedload in these experiments. The differing substrates create different fluvial environments for which we use decorrelation analysis to measure a channel time scale. In addition, the difference in color between the red or blue dye-filled water and the white, black, brown, or green substrate allows the separation of wet and dry pixels by the method indicated in Section 2.2.3.

The following sections in this chapter are devoted to describing the parameters and boundary conditions of the analog experiments, with respect to the relationships that we draw from them.

5.2. Sediment Flux, Aggradation, and Froude Number

In a simple system of sand and water, water carves out channels and sand fills them back in. Therefore, we hypothesized that the channel-mobility time scale should scale relative to the proportions of sand and water. The particular simple systems chosen for this analysis were two subsequent experiments performed in the delta basin at the Saint Anthony Falls Laboratory of the University of Minnesota, DB03-1 and DB03-2. Both of these experimental deltas were grown to the desired size, after which sea-level rise was applied to match aggradation (slightly inaccurately in DB03-2: see later description) such that the surface area of an aerial view of the delta would remain constant with time. The images used for analysis came from this steady-state aggradational phase, after the delta had reached equilibrium with its new conditions.
The high-resolution topography experiment (DB03-1) (Sheets et al, in press, 2007) provides a standard experimental deltaic setting. The sediment consisted of 70% quartz sand (0.120 mm) and 30% bimodal anthracite (0.190 and 0.460 mm) to serve as fines (see Table 5.1), and entered the system from a single feed which sent in pre-mixed sand and water. The ratio of volumetric water flux to sediment flux, $Q_w:Q_s$, was 40:1. The average aggradation rate of this experiment was 5.0 mm/hr, and the average self-built slope ($S$) of the fan was 0.05. The subaerial surface of the fan was a quarter-circle of radius 2.5 meters, which stayed constant throughout the experiment. (All of the above data are from the work of Sheets et al., in press, 2007.) Pseudo-subsidence from base-level rise was set to balance aggradation. The Froude number of this system is estimated by

$$Fr = \sqrt{\frac{S}{c_f}}$$  \hspace{1cm} (5.1)

(Martin et al., manuscript in preparation, 2007), where $c_f$ is the friction coefficient, which is estimated to be 0.01 (Martin et al., manuscript in preparation, 2007). Substituting this value for $c_f$ gives a direct relationship between Froude number and slope:

$$Fr = 10\sqrt{S}$$  \hspace{1cm} (5.2)

Solving for a slope of 0.05 shows that the Froude number in the high-resolution experiment was approximately 2.2, which places it in the supercritical field.
The low-Froude-number experiment (DB03-2) had a ratio of volumetric water flux to sediment flux of approximately 10,000:1 for the sequences of the experiment that were analyzed for channel migration. This value approaches the value seen in natural rivers (Martin et al., manuscript in preparation, 2007). The sediment in this experiment consisted of 70% quartz sand (0.12 mm) and 30% bimodal anthracite (70% fine, 0.10 mm, and 30% coarse, 0.50 mm) to serve as the “fines” in the system (Table 6.1). The aggradation rate was 0.030 mm/hr, which was balanced by 0.030 mm/hr of sea-level rise as pseudo-subsidence. The subaerial fan area was a quarter-circle of between approximately 2.65 and 2.8 meters: in this experiment, base-level rise almost (but not quite) matched incoming sediment flux. Measured deltaic slope averaged 0.009 and ranged between 0.001 and 0.013. Measured Froude numbers averaged at 0.63 and ranged between 0.29 and 1.58. (All data in this paragraph are from the work in Martin et al., manuscript in preparation, 2007.) The theoretical Froude number obtained by Equation 6.2 and using the average slope of 0.009 is 0.95. This is higher than the measured Froude number, but not so much higher that a recalibrated value of $c_f$ in Equation 6.1 would result in a subcritical state during DB03-1.
Figure 5.1. Photo of DB03-1. This photo of the experiment is taken from the data of Ben Sheets, as presented in Sheets et al. (in press, 2007).

Figure 5.2. Photo of DB03-2. This photo of the experiment is taken from the data of John Martin, as presented in Martin et al. (manuscript in preparation, 2007).
Table 5.2. Experimental parameters for the low-Froude-number and high-resolution experiments.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>$Q_w$ (L/s)</th>
<th>$Q_s$ (L/s)</th>
<th>$Q_w$:$Q_s$</th>
<th>$((dη/dt)_{avg})$ (mm/hr)</th>
<th>$S_{avg}$</th>
<th>&lt;Fr&gt; calc’d</th>
<th>&lt;Fr&gt; measured</th>
</tr>
</thead>
<tbody>
<tr>
<td>DB03-1 (High-Res)</td>
<td>0.400</td>
<td>0.0100</td>
<td>40:1</td>
<td>5.000</td>
<td>0.05</td>
<td>2.2</td>
<td>-</td>
</tr>
<tr>
<td>DB03-2 (Low&lt;Fr&gt;)</td>
<td>0.400</td>
<td>-0.000040</td>
<td>~10000:1</td>
<td>0.023</td>
<td>0.009</td>
<td>0.95</td>
<td>0.63</td>
</tr>
</tbody>
</table>

$Q_w$ is total measured water flux, $Q_s$ is total measured sediment flux, $((dη/dt)_{avg})$ is average measured aggradation rate, $S_{avg}$ is average measured slope, <Fr> calc’d is calculated Froude number, and <Fr> measured is measured Froude number.

Note that, in DB03-2, there are discrepancies between the measured sediment flux, the measured aggradation and the aggradation from what the Exner equation would lead one to expect. This is because the sediment feeder was pushed to its absolute minimum in terms of sediment supply, so small fluctuations occurred that allowed sediment feed to be, on average, a little higher than its measured value.

5.3. Cohesiveness of Sediment and Base Level

The goal of this analysis was to discern the channel mobility within an experimental delta in which base level was first held constant, and then set to rise in order to balance incoming sediment flux, such that the subaerial surface of the delta retained a constant area. This delta experiment was performed in the delta basin at the Saint Anthony Falls Laboratory and used the new cohesive sediment mixture developed by Hoyal and Sheets (manuscript in preparation, 2007).

The setup for this experiment was the same as for the low-Froude-number and high-resolution experiments, except in the sediment type being used, and the factors listed in table 5.3.
Table 5.3. Experimental parameters for the cohesive sediment experiment.

<table>
<thead>
<tr>
<th>Cohesive Sediment: Phase</th>
<th>( Q_w ) (L/s)</th>
<th>( Q_s ) (L/s)</th>
<th>( Q_w:Q_s )</th>
<th>((dn/dt)_{avg}) (mm/hr)</th>
<th>&lt;Fr&gt;</th>
</tr>
</thead>
<tbody>
<tr>
<td>Base-Level Constant</td>
<td>0.1667</td>
<td>0.00017</td>
<td>980.6:1</td>
<td>0</td>
<td>~1</td>
</tr>
<tr>
<td>Base-Level Rise</td>
<td>0.1667</td>
<td>0.00017</td>
<td>980.6:1</td>
<td>0.237</td>
<td>~1</td>
</tr>
</tbody>
</table>

The parameters listed here are defined as in Table 5.2 and in the “list of terms” at the end of this chapter.

Figure 5.3. Photo of the 2007 cohesive sediment experiment. Photo is taken at an angle: the sediment and water feed is in the lower-left-hand corner and the delta is building into the ocean in the upper right corner. (Photo and Experiment by John Martin. Cohesive-sediment mixture by Hoyal and Sheets, in press, 2007.)
5.4. Base Level

Rise or fall in base level causes changes in deltaic stratigraphy, as well as changes in the channel patterns on the surface of the delta. A major goal of the 2002 run of the Experimenta EarthScape Facility (XES02) was to understand the dynamics of a deltaic system undergoing eustatic change, and the characteristics of the deposits left from eustatic cycles. In particular, slow, fast, and superimposed slow and fast base-level cycles were imposed on the delta, which was also undergoing simulated passive-margin subsidence on a linear hinge, occurring at a constant rate. This linear hinge extended for the upstream 4.2 meters of the basin. At the upstream end, subsidence was 0 mm/hr, and at the downstream end, subsidence was 3.71 mm/hr (Kim et al., 2006). The last 1.8 meters of the basin subsided at a constant rate of 3.71 mm/hr. The following diagram shows sea level in the experiment:

![Diagram of sea level with time for XES02 run](image)

**Figure 5.4.** (from Kim, 2007, Figure 3.2) Sea level with time for XES02. Data were taken during the initial stillstand (10-18 hours), halfway through the base-level fall and base-level rise of the slow cycle, and for the full duration of the rapid cycle.
We applied the decorrelation analysis to the early stillstand (hours 10 to 18), the slow cycle (hours 26 to 134, which is a duration of 108 hours), and the rapid cycle (hours 144 to 162, which is a duration of 18 hours). Data were taken from the middle of the base-level fall and base-level rise sequences of the slow cycle (in order for the sinusoidal sea-level change to be approximately linear during data collection), and from the entire duration of the rapid cycle. In both of these cycles, base level fell 11 cm and then rose to its previous depth (Kim, 2007).

A single point source fed a pre-mixed combination of both water and sediment. Water entered the system at a rate of 0.417 L/s, and sediment entered the system at 0.00506 L/hr. The ratio of volumetric water flux to volumetric sediment flux was therefore 82.4:1. This sediment was composed of 63% quartz sand (0.110 mm), 27% anthracite coal (0.460 mm and 0.190 mm) to serve as the “fines” because of its low specific gravity (1.3), and 10% kaolinite to improve the stability of the deposit by adding some cohesion. (All information on experimental parameters are from the work of Kim et al., 2006, and Kim, 2007.)

**Table 5.4.** Experimental parameters for XES02.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>$Q_w$ (L/s)</th>
<th>$Q_s$ (L/s)</th>
<th>$Q_w : Q_s$</th>
<th>$((d\eta/dt)_{avg})$ (mm/hr)</th>
<th>$S_{avg}$</th>
<th>$&lt;Fr&gt;$ calc’d</th>
<th>$&lt;Fr&gt;$ measured</th>
</tr>
</thead>
<tbody>
<tr>
<td>XES02</td>
<td>0.417</td>
<td>0.00516</td>
<td>82.4:1</td>
<td></td>
<td></td>
<td></td>
<td>-</td>
</tr>
</tbody>
</table>

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Figure 5.5. Photos showing base-level changes in XES02.

The picture showing base-level fall (top) shows the channel in incised valley that it created for itself. The semicircular shape extending towards the ocean from is the large, stable delta lobe created during this phase of the experiment extending out of the incised valley.

The picture showing base-level fall (bottom) shows several migrating bars in a wide, braided channel that is rapidly aggrading as base-level rise floods the fan.

(Images taken from the data of Wonsuck Kim (Kim et al., 2006; Kim, 2007).)
Figure 5.6. (from Kim, 2007, Figure 2.1) A diagnostic sketch of the experimental set-up for XES02.
5.5. Vegetation

It is a well-known fact that vegetation slows the migration of river channels (Tal et al., 2006, and references therein; Tal and Paola, 2007, and references therein; Simon and Collison, 2000). Quantifying the effects of vegetation on channel migration and avulsion can be difficult in the field, so an experimental analog provides a good setting to build a better understanding of the effects of vegetation on channel mobility.

The braided and vegetated dynamics experiment was run in a flume that was 16 meters long by 2 meters wide, with a slope of 0.015 and well-sorted pure quartz sand with a $D_{50}$ of 0.5 mm. Roughness elements were placed at the sides of the flume to diminish sidewall effects. The data that we analyzed came from the 2002 run, in which there was one sediment and water feed at the upstream end of the channel. In the first stage of the experiment, water flux was set at bankfull to overbank discharge (2.0 L/s) and a steady-state braided channel developed in which sediment output equalled sediment input, which was 3.5 g/s. After this, $Q_w$ was reduced to 0.40 L/s, and the flume was seeded with alfalfa at a density of 1 seed per cm$^2$. This value for the water flux left the channels too shallow to be able to transport sediment, and allowed the alfalfa time to take hold and grow. Therefore, no sediment was added to the system, because there was almost no sediment motion and it was desired to keep the sediment budget in steady state. This low-flow period of vegetated entrenchment lasted six days. Then, water flux was increased to 2.0 L/s, which was bankfull to overbank discharge. Flow during floods was turbulent. Sediment flux during the floods was 3.5 g/s, the same as during the non-vegetated phase, and matched the sediment output rate. Flow velocities during the floods were in the range
of 0.2 to 0.3 meters per second, and flow depths were between 1 and 3 centimeters. This provided a range of Froude numbers \( (u/(gh)^{0.5}) \) between 0.37 and 0.96, which are in the subcritical range, but the average Froude number was usually on the high end of this range, and passed into supercritical flow. Each of these floods lasted two hours. After every two hours, discharge rates were lowered again to 0.40 L/s, the flume was reseeded at the same density of 1 seed per cm\(^2\), and the low flow persisted again for six days, after which another 2-hour flood was run. This pattern continued for a total of seventeen floods. (All of the data presented in this paragraph are from the work of Tal and Paola, 2007.)

**Table 5.5.** Fluxes in the braided and vegetated dynamics experiment

<table>
<thead>
<tr>
<th>Phase of Braid/Veg</th>
<th>( Q_w ) (L/s)</th>
<th>( Q_{s_{in/out}} ) (L/s)</th>
<th>( Q_w/(Q_i) )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nonvegetated</td>
<td>2.0</td>
<td>0.0020</td>
<td>1000:1</td>
</tr>
<tr>
<td>Vegetated; Flood Stage</td>
<td>2.0</td>
<td>0.0020</td>
<td>1000:1</td>
</tr>
<tr>
<td>Vegetated; Low Flow</td>
<td>0.40</td>
<td>0</td>
<td>-</td>
</tr>
</tbody>
</table>

**Table 5.6.** Flood timing

<table>
<thead>
<tr>
<th>Phase of Braid/Veg</th>
<th>Flood Duration ( (Dur_{flood}) )</th>
<th>Intermittency ( (I) )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nonvegetated</td>
<td>Full Runtime</td>
<td>0</td>
</tr>
<tr>
<td>Vegetated</td>
<td>2 hours</td>
<td>6 days</td>
</tr>
</tbody>
</table>
Figure 5.7. **Photos of the Braided and Vegetated Dynamics Experiment.** In these photos, pink is water (dyed with rhodamine), white is quartz sand, and green is alfalfa (*Medicago sativa*), which served as simulated vegetation.

(continued above)

The top two photos are perspective views of the flume in which the braided and vegetated dynamics experiment was run. (top left) The non-vegetated stage of the experiment had shallow, unconstrained flow and was highly braided. (top right) In the vegetated stage of the experiment, simulated vegetation (alfalfa) constrained the flow into a few main channels and greatly reduced rates of migration. Note the existence of point-bars and cut-banks, including a cutoff in an element of sinuosity.

The bottom two photos are stitched and ortho-rectified panoramas of the map view of the experimental channels. (bottom upper) Vegetation has forced the fluvial system into a few channels. Notice the elements of sinuosity near the center of the image, including the point-bars, cut-banks, and cutoff. (bottom lower) The nonvegetated and fully braided stage of this experiment consists of freely flowing water that is constantly reshaping the various sandbars situated across the flume.

(continued below)
List of Terms

$S$ = Slope

c$_f$ = Dimensionless friction coefficient, $\approx 0.01$

Fr = Froude number, $= \left(\frac{u}{gh^{0.5}}\right)$

$u$ = Flow velocity

$g$ = Acceleration due to gravity

$h$ = Flow depth

$Q_w$ = Volumetric water flux (L/s)

$Q_s$ = Volumetric sediment flux (L/s); volume given includes grains and pore spaces

$dn/dt$ = Aggradation rate (not including subsidence: just from sediment deposition)

$S_{avg}$ = Average deltaic surface slope

$Q_{in/out}$ = Volumetric sediment flux both entering and leaving the system. This volumetric
sediment flux includes the volume of pore spaces in-between grains.

$D_{50}$ = Median grain size

$Dur_{flood}$ = Flood duration, for the braided and vegetated dynamics experiment. This
equals 2 hours.

$I$ = Intermittency: the amount of time that passes between floods. This equals 6 days for
the braided and vegetated dynamics experiment.
Chapter 6: Results and Analysis

6.1. Sediment Flux, Aggradation, and Froude Number

6.1.1. Results

Two experiments, DB03-1 and DB03-2 (Section 5.2.1), were compared in order to understand the impact of sediment flux, aggradation, and Froude number on the channel migration in a deltaic system. (In actuality, because the rate of sea-level rise was set to equal the mass of sediment entering the system, sediment flux and aggradation scale approximately linearly: base-level rise in DB03-2 that was slightly too slow causes small discrepancies from an absolute statement of proportionality between sediment flux and aggradation.) The first four of the following plots show the randomness-scaled correlation of a single experiment. The first two are of data sets taken from the low-Froude-number experiment (from approximately 830-1580 hours and 1800-2400 hours in runtime), and the second two are data sets taken from the high-resolution experiment.

We fitted two curves to each of these four data sets. The first curve is the best-fit exponential decay for that particular data set. The second curve is the exponential decay in which the value for \( a \) is the average value for \( a \) across all four data sets under analysis and the value for \( c \) is zero, which is close to the average value for \( c \) (-0.02221), and produces graphs with a higher \( R^2 \). We believe these averages to be analytically valid because the values for \( a \) and \( c \) between these experiments are generally very close. This
second curve is used to compare the decay constants from each of the experiments and determine the relative rates of channel mobility in the various experimental systems.

The final two plots (Figures 6.5 and 6.6) compare the “for comparison” curves of all four experimental settings, in order to graphically display the difference in channel mobility between DB03-1 and DB03-2. Correlation is scaled to randomness, and all of the times given are in hours; this convention holds true for every analysis in this thesis.

**Figure 6.1.** Data and their fitting exponential decay curves: hours 830-1580 of the low-Froude-number experiment (DB03-2).

<table>
<thead>
<tr>
<th></th>
<th>Best Fit</th>
<th>Fit For Comparison</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>0.5774</td>
<td>0.6602</td>
</tr>
<tr>
<td>M</td>
<td>0.03288</td>
<td>0.02151</td>
</tr>
<tr>
<td>c</td>
<td>0.03288</td>
<td>0</td>
</tr>
<tr>
<td>$R^2$</td>
<td>0.64</td>
<td>0.54</td>
</tr>
</tbody>
</table>

**Table 6.1.** Correlation with time: DB03-2, hrs. 830-1580.

**Figure 6.2.** Data and their fitting exponential decay curves: hours 830-1580 of the low-Froude-number experiment (DB03-2).

<table>
<thead>
<tr>
<th></th>
<th>Best Fit</th>
<th>Fit For Comparison</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>0.7305</td>
<td>0.6602</td>
</tr>
<tr>
<td>M</td>
<td>0.04589</td>
<td>0.03015</td>
</tr>
<tr>
<td>C</td>
<td>0.05558</td>
<td>0</td>
</tr>
<tr>
<td>$R^2$</td>
<td>0.65</td>
<td>0.57</td>
</tr>
</tbody>
</table>

**Table 6.2.** Correlation with time: DB03-2, hrs. 830-1580.
Figure 6.3. Data and their fitting exponential decay curves: 01/23 run of the high-resolution topography experiment (DB03-1).

Table 6.3. Correlation with time: DB03-1, 01/23 run.

<table>
<thead>
<tr>
<th></th>
<th>Best Fit</th>
<th>Fit For Comparison</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>0.7185</td>
<td>0.6602</td>
</tr>
<tr>
<td>M</td>
<td>1.515</td>
<td>1.782</td>
</tr>
<tr>
<td>c</td>
<td>-0.04842</td>
<td>0</td>
</tr>
<tr>
<td>$R^2$</td>
<td>0.40</td>
<td>0.38</td>
</tr>
</tbody>
</table>

Figure 6.4. Data and their fitting exponential decay curves: 02/04 run of the high-resolution topography experiment (DB03-1).

Table 6.4. Correlation with time: DB03-1, 02/04 run.

<table>
<thead>
<tr>
<th></th>
<th>Best Fit</th>
<th>Fit For Comparison</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>0.6264</td>
<td>0.6602</td>
</tr>
<tr>
<td>M</td>
<td>0.2788</td>
<td>0.8465</td>
</tr>
<tr>
<td>c</td>
<td>-0.1721</td>
<td>0</td>
</tr>
<tr>
<td>$R^2$</td>
<td>0.50</td>
<td>0.46</td>
</tr>
</tbody>
</table>

Figure 6.5. Correlation vs. time: DB03-1 and DB03-2. Linear axes. $a = 0.6602; c = 0; M =$ (see Table 6.5)

Table 6.5. Correlation with time: DB03-1 and DB03-2.

<table>
<thead>
<tr>
<th>Experimental Run</th>
<th>M</th>
</tr>
</thead>
<tbody>
<tr>
<td>HR 01/23</td>
<td>1.782</td>
</tr>
<tr>
<td>HR 02/04</td>
<td>0.8465</td>
</tr>
<tr>
<td>LF 830-1580 hrs</td>
<td>0.02151</td>
</tr>
<tr>
<td>LF 1800-2400 hrs</td>
<td>0.03015</td>
</tr>
</tbody>
</table>
6.1.2. Analysis

6.1.2.a. Balance of Sediment and Base-Level Rise

There exists a special case in deltaic stratigraphy, often seen in experiments, such that incoming volumetric sediment flux is in balance with sea-level rise and/or subsidence. This causes the subaerial surface area of the delta to remain constant. (In DB03-1, sea-level rise is set exactly to equal the amount of delta building from deposition. In DB03-2, this is almost true, but its shoreline of the delta progrades at a rate of 0.09 mm/hr.)

Therefore, these ideal deltas always have the same subaerial surface area. This means that all of the sediment is being deposited on the delta, and that volumetric sediment flux into the system, divided by fan surface area, is the aggradation rate: deposition scales linearly with sediment flux. This can be demonstrated by the Exner erosion equation:

$$\frac{\partial \eta}{\partial t} = -\frac{1}{\epsilon_0} \nabla \cdot Q_s$$  \hspace{1cm} (6.1)
In the Exner equation, \( \eta \) is surface elevation with respect to an internal datum (i.e. it is entirely a sedimentological term, ignoring subsidence and/or uplift), \( \varepsilon_o \) is 1 minus the porosity of the sediment, and \( Q_s \) is total sediment flux. In order to simulate the experimental conditions, we set all of the incoming sediment flux \( (Q_{s,in}) \) to deposit on the delta surface.

\[
Q_{s,in} = -\nabla \cdot Q_s \tag{6.2}
\]

\[
Q_{s,out} = 0 \tag{6.3}
\]

\( Q_{s,out} \) is the total volumetric sediment flux that passes off the deltaic system and into the ocean basin: this is the part of sediment flux that causes the delta to aggrade. Now, we are able to substitute our findings back into the Exner equation. Remembering that we are using volumetric sediment flux, in which the given volume includes the pore spaces (thus canceling out the \( 1/\varepsilon_o \) term), the relationship between incoming volumetric sediment flux and aggradation is given by equation 6.4:

\[
\frac{\partial \eta}{\partial t} = \frac{Q_{s,in}}{A_{\text{delta}}} \tag{6.4}
\]

where \( A_{\text{delta}} \) is the surface area of the delta.
6.1.2.b. Scaling DB03-1 and DB03-2 to Each Other

In order to compare these two delta basin experiments, we first need to scale their attributes. These two experiments are alike, in all but five ways. First, DB03-1 had a much higher sediment flux, and therefore much higher rate of base-level rise ("pseudo-subsidence") than DB03-2. Second, the flow in DB03-1 was supercritical (Froude number > 1), while the flow in DB03-2 was subcritical (Froude number < 1).

Third, the slope of DB03-2 is significantly lower than the slope of DB03-1, though it is higher than is expected for the given sediment flux. In fact, the a ratio of the surface slope to sediment flux in DB03-2 is approximately 40 times that for DB03-1. This is due to the tendency of flow to spread laterally, cover a greater area of the fan, and become less efficient at transporting sediment as fluvial slope decreases. This can be seen in the difference of wetted areas between DB03-1 and DB03-2. In DB03-1, 34% of the surface was wetted, on average. In DB03-2, the average wetted area was 55%: the flow had become wider and shallower. (A more thorough explanation of the ideas presented in this paragraph is given in Martin et al. (manuscript in preparation, 2007).)

The fourth difference is trickier to deal with: in DB03-2, he fan area of increases by approximately 15 cm throughout the part of the experimental run that we analyzed, from approximately 2.65 meters to 2.8 meters. In comparison, the radius of DB03-1 stayed constant at 2.5 meters. Sea-level rise did not exactly equal aggradation. This requires us to think of two possible effects: the effect of the size of the fan on the rate at which
channels will be filled, and the fact that not all of the sediment is being used to fill channels on the surface of the fan: some of it is being deposited offshore.

Fortunately, these do not have large effects. The results from XES02 show us that changes in base level which are slow with respect to a channel time scale do not cause statistically significant changes in channel mobility. In fact, data from DB03-2 show us that the discrepancies that are caused by the prograding shoreline are smaller than the error in measuring the sediment flux. Therefore, we will treat DB03-2 as if it were in perfect volume balance (such that the principles stated in Section 6.1.2.a apply to it), though keeping in the back of our minds that it is not.

The increased size of the delta presents a problem of channel volumes. According to the relations developed by Parker et al. (in press, 2007) for gravel-bed rivers, water flux scales linearly with channel cross-sectional area. A common-sense approach of balancing areas shows that this should be true for fluvial systems in general. DB03-1 and DB03-2 had the same water flux, but DB03-2 was longer. Therefore, full channel volume should be greater in DB03-2 by a factor that is the ratio of the deltaic lengths, which is 6% to 12%. This means that DB03-2 would need to fill 6-12 percent more channel volume in order to reach the same amount of decorrelation as DB03-1.

However, these experiments also contain overbank flow, which is not channel-forming and is therefore not considered in the relations of Parker et al. (in press, 2007). In particular, DB03-2 has more overbank flow than DB03-1 (its wetted area increases by
We believe that this effect is significant, and therefore expect that the channel volumes of DB03-2 were either the same as or less than those in DB03-1.

6.1.2.c. Channel Mobility and Deposition

The ratio of the decay constants show that the channels in DB03-1 moved at 51 times the rate of channels in DB03-2. Both the sediment flux and the aggradation rate of DB03-1 were approximately 213 times the sediment flux and aggradation rate of DB03-2. We seek to apply a simple channel-filling argument to this system, to see if aggradation is the only major factor causing these channels to switch. Therefore, we scaled the average of the two calculated decay constants for DB03-1 and DB03-2 (denoted $M_{HR}$ and $M_{LF}$, respectively) to the aggradation rates of their respective experiments. (Note that mobility is in units of 1/hours.) We assumed offshore deposition to be approximately equal to zero in both of these experiments, thus allowing this scaling to work. The resulting expression for aggradation-scaled mobility ($M_{Agg}$) is

$$M_{Agg} = \frac{M}{\frac{d\eta}{dt}}$$

where $d\eta/dt$ is aggradation and $M$ is the mobility of a particular experimental system (here we define it as the average of our measurements for $M$ for each system). Table 6.1 gives the values for channel mobility between the two experiments in both the scaled and unscaled cases.
Table 6.6. Channel mobility scaled to aggradation in DB03-1 and DB03-2.

<table>
<thead>
<tr>
<th></th>
<th>M (1/hr)</th>
<th>$M_{Agg}$ (1/mm)</th>
<th>$d\eta/dt$ (mm/hr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>High-Res (DB03-1)</td>
<td>1.3143</td>
<td>0.2629</td>
<td>5.000</td>
</tr>
<tr>
<td>Low-&lt;Fr&gt; (DB03-2)</td>
<td>0.02583</td>
<td>1.1230</td>
<td>0.023</td>
</tr>
</tbody>
</table>

Table 6.2 shows the ratios of the channel mobilities, both scaled and non-scaled. If the relationship between mobility and sediment flux were linear, mobility scaled to sediment flux should always be constant, and the ratio of the mobilities would be one. Instead, the ratio of scaled channel mobilities shows that, scaled to aggradation (or sediment flux), the channels in the low-Froude-number experiment are moving at approximately four times the rate of the channels in the high-resolution experiment.

Table 6.7. Ratios between the unscaled mobilities and the aggradation-scaled mobilities for DB03-1 (HR) and DB03-2 (LF).

<table>
<thead>
<tr>
<th></th>
<th>$M_{HR}$ / $M_{LF}$</th>
<th>$M_{Agg,HR}$ / $M_{Agg,LF}$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>50.9</td>
<td>0.234</td>
</tr>
</tbody>
</table>

Note that the experiment (HR) with the higher sediment flux has a much higher unscaled mobility, but when scaled by aggradation, the mobility is actually greater in the experiment with the lower sediment flux (LF).

This nonlinearity raises questions about the simple hypothesized argument of the "fillability" of the channel. If it is truly deposition that is controlling migration and
avulsion, as it should seem to be in this simple system of sand and downhill-flowing water, why doesn’t channel migration scale linearly to volumetric sediment flux? In order to answer this question, we look to the effects of the Froude number and slope on the channel mobility in this system.

We hypothesize that a large part of the nonlinearity was caused by the presence of ripples in the low-Froude-number experiment. Direct observations of these ripples show that they were large, with heights on the order of one full channel depth. Because of this, they behaved more as channel bars than as ripples, and acted to re-route the flow (Martin et al., manuscript in preparation, 2007). This large ripple size is easy to understand. Ripples, in fact, have approximately the same size in a very wide range of natural water flows. Grazer (1982) showed that, by Reynolds-Froude scaling, the length scale of the ripples goes as the kinematic viscosity to the $2/3$ power. DB03-2 was run at standard cold tap water temperatures, which is comparable to or slightly colder than standard river temperatures. Because water viscosity varies with the size of the ripples formed in the experiment would be the same size as those formed in standard river channels. However, because the channels in DB03-2 were approximately 1 centimeter deep, ripple heights were constrained to flow depth, and these flow-depth-high ripples they increased channel mobility significantly by diverting the flow.

Therefore, Martin et al. (manuscript in preparation, 2007) explains that, although a subcritical experiment could be expected to be beneficial in its ability to better simulate natural systems (most analog experiments of deltaic environments are run with Froude
number greater than one), difficulties in Froude scaling make a subcritical model create more scaling problems than it solves. However, this does not totally invalidate the usefulness of a subcritical model, inasmuch as these ripples seem act in the same way as channel bars do in large rivers.

A second possible reason for the faster channel mobility in DB03-2 than DB01-1, relative to aggradation, is the deltaic slope. Low deltaic slope caused DB03-2 to be covered in wide, shallow flow, which is naturally more prone to depositing sediment and migrating than deep, narrow flow.

6.2. Cohesiveness of Sediment and Base Level

6.2.1. Results

Curves of correlation with time were measured for the base-level constant and base-level rise phases of the 2007 cohesive sediment experiment, and are reproduced in Figure 6.7. (Base-level rise is set to match volume of incoming sediment, such that the subaerial surface area of the fan remains constant.) The value for the decay constant for the phase with base level held constant is 0.1342, and is 0.2081 for the phase in which base-level rises.

This is the only experiment in which the author did not perform all of the analysis. John Martin prepared these images in the same way as all of the others, and also despeckled
them and removed bodies of standing water. The following plot shows the curves of randomness-scaled correlation for the 2007 cohesive sediment experiment. As in the other experiments, the values of \( a \) and \( c \) were normalized to their average value for the two data sets.

6.2.2. Analysis

The ratio of the mobilities for the base-level constant and base-level rise phases of the cohesive sediment experiment are

\[
\frac{M_{BL\text{const}}}{M_{BL\text{rise}}} = \frac{0.1342}{0.2081} = 0.6449
\]  

(6.8)

where \( M_{BL\text{const}} \) is channel mobility (the decay constant for correlation against time) in the base-level constant phase of the experiment, and \( M_{BL\text{rise}} \) is channel mobility in the base-level rise phase of the experiment. Equation 6.8 shows that the channels in the base-level rise phase of the cohesive delta experiment were mobile at 1.55 times the rate of rate of the channels in the base-level constant phase. We expect that base-level rise caused more deposition on the subaerial delta surface. Increased deposition on deltaic lobes decreases lobe time scales, and aggradation within the main channels causes them to superelevate and become more avulsive.

We compare three characteristic ratios in order to quantify channel mobility with respect to base-level change. First, we use Equation 6.8 for relative mobility (1.55) with respect to base-level rise (0.237 mm/hr). This ratio is 6.3. Then, we do the same comparison,
except with additional mobility (0.55x), and the characteristic ratio is 2.3. Third, we compare approximate average stillstand channel mobility (0.13/hour) to base-level rise as a characteristic ratio with which we can compare the effects on channel mobility of base-level rise with respect to average channel mobility. The resulting characteristic ratio is 0.55/mm. We do not have enough data here to perform a full analysis, but we use this third characteristic ratio in section 6.3 in comparison with results from XES02 to look at a quantification of channel mobility with respect to base-level changes.

Figure 6.7. Correlation with time: delta in cohesive sediment: base-level rise vs. base-level constant. The equations provided on the plot are for the normalized $a$ and $c$ values. The original curves for these two equations are given in table 6.3.
Table 6.8. Constants in the exponential decay curves for the base-level rise and base-level constant cases of the 2007 cohesive sediment experiment.

<table>
<thead>
<tr>
<th></th>
<th>a</th>
<th>M</th>
<th>c</th>
</tr>
</thead>
<tbody>
<tr>
<td>base-level rise</td>
<td>0.6872</td>
<td>0.2307</td>
<td>0.1452</td>
</tr>
<tr>
<td>base-level constant</td>
<td>0.6136</td>
<td>0.1209</td>
<td>0.1379</td>
</tr>
</tbody>
</table>

It is important to note that there is considerable scatter in the data, and that these relationships are averaged (Figure 6.7). We believe that this is related the low degree of allogenic forcing in this experiment. In XES02, no significant change in mobility occurred during the base-level cycle that was slow with respect to a channel-mobility time scale. However, significant channel change occurred when base level varied at a rate that was much faster than a channel time scale. The 2007 cohesive-sediment experiment provides us with an intermediate case, in which signal of increased channel mobility during base-level rise is strong enough to appear but is weak enough that it appears only when we are able to average across all of the data.

6.3. Base-Level Change

6.3.1. Results

Our analysis of XES02 measured channel response to base-level change. The plots are not given here, to save space, but the general equations for the different sequences of the experiment are given as follows in Table 6.9. When normalized such that $a$ and $c$ are equal across the board, the relations given in Table 6.10 emerge.
Table 6.9. Coefficients for the exponential decay in XESO2.

<table>
<thead>
<tr>
<th>Fit ( ( a(e^{-M*t}+c) ) )</th>
<th>a</th>
<th>M</th>
<th>c</th>
<th>( R^2 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Steady*</td>
<td>0.6914</td>
<td>0.7766</td>
<td>0.1152</td>
<td>0.67</td>
</tr>
<tr>
<td>Slow fall</td>
<td>0.6902</td>
<td>3.872</td>
<td>0.1149</td>
<td>0.54</td>
</tr>
<tr>
<td>Slow rise</td>
<td>0.5593</td>
<td>1.703</td>
<td>0.0521</td>
<td>0.44</td>
</tr>
<tr>
<td>Fast Fall (Incised Valley Portion)</td>
<td>0.5369</td>
<td>2.094</td>
<td>0.2717</td>
<td>0.32</td>
</tr>
<tr>
<td>Fast Rise (part 1)*</td>
<td>0.6001</td>
<td>7.282</td>
<td>0.2895</td>
<td>0.40</td>
</tr>
<tr>
<td>Fast Rise (part 2)*</td>
<td>0.4946</td>
<td>5.12</td>
<td>0.4183</td>
<td>0.38</td>
</tr>
</tbody>
</table>

* We do not know what caused this mobility to be so low. We weakly hypothesize that, since this occurred at the beginning of the experiment, the delta was not responding to any previous conditions.

Table 6.10. Coefficients for the exponential decay in XESO2, with \( a \) and \( c \) normalized.

<table>
<thead>
<tr>
<th>Fast Cycle Normalized (without part 2 of base-level rise)*</th>
<th>a</th>
<th>M</th>
<th>c</th>
<th>( R^2 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fast Fall (Incised Valley Portion)</td>
<td>0.5685</td>
<td>6.304</td>
<td>0.2806</td>
<td>0.32</td>
</tr>
<tr>
<td>Fast Rise (part 1)*</td>
<td>0.5685</td>
<td>2.406</td>
<td>0.2806</td>
<td>0.4</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Slow Cycle Normalized</th>
<th>a</th>
<th>M</th>
<th>c</th>
<th>( R^2 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Slow fall</td>
<td>0.62475</td>
<td>2.752</td>
<td>0.083515</td>
<td>0.51</td>
</tr>
<tr>
<td>Slow rise</td>
<td>0.62475</td>
<td>2.503</td>
<td>0.083515</td>
<td>0.42</td>
</tr>
</tbody>
</table>

Note that we used only one part of the analysis from the base-level rise sequence, because it is imperative that the values for \( a \) and \( c \) be close before averaging, in order to ensure analytical rigidity.

6.3.2. Analysis

These relationships show that the slow base-level cycle had little statistically significant effect on channel mobility. This makes sense, in that the slow cycle was designed such that the change in base level occurred on a long time scale with respect to a channel mobility time scale.

The channels in the base-level rise segment of the rapid cycle moved at approximately three times the rate of the channels in the base-level fall sequence. This is because the channels incised and entrenched themselves during the base-level fall cycle, and
prograded outwards with a large, stable deltaic lobe. During the base-level rise sequence, deposition on the shoreline caused rapid mouth-bar migration, increased rates of lobe over-building, and increased rates of main channel sedimentation. We believe that this caused unfavorable water surface slopes (consistent with the ideas presented in Slingerland and Smith, 1998, 2004), and increased main-channel superelevation (consistent with the ideas of Bryant et al., 1995; Heller and Paola, 1996; and Mohrig et al., 2000). This combination of unfavorable slope and superelevation led to increased observed channel mobility due to the avulsion and relocation of the main channel. We also believe that channel mobility also increased due to the more rapid mouth-bar migration in response to base-level rise. (As a side note, the values for the “confinement c” are high in the base-level rise and fall, because after the valley was cut, the fluvial action in both sequences was trapped inside it.)

We are interested in comparing rates of base-level change with rates of channel migration and avulsion. During the slow cycle, average rate of base-level change was 2.0 mm/hr. During the rapid cycle, average rate of base-level change was 12.2 mm/hr. This means that a six-fold change in rate of base-level change resulted in a three-fold change in channel mobility. If we take a characteristic channel mobility to be approximately 3/hour (an approximate average of the given channel mobilities), then this tells us that a ratio of average channel mobility to base-level change of gives a characteristic ratio of approximately (1/3)/mm for the fast cycle and approximately 2/mm for the slow cycle. We can compare this with the value 0.55/mm for the cohesive delta, which showed a 50% increase in channel mobility from the stillstand when base-level rise was initiated.
We believe that these values and mobilities could show a linear relationship between base-level and mobility, with respect to “0” mobility defined to be mobility at stillstand. However, we would need more data in order to make any conclusive quantitative statement. At the very least, these show that the rate of base-level change with respect to channel timescale could be important in determining channel mobility. In the future, we would like to do additional work to quantify the relationship between base-level and channel mobility more rigorously, using a framework like the one we present here.

6.4. Vegetation

6.4.1. Results

Each flood in the vegetated phase of the braided and vegetated dynamics experiment was too short to do a full decorrelation analysis. In addition, decorrelation during low flow and effects from moving the dye trays (used to calibrate channel depth; see Tal and Paola, 2007) cause some distortion to the overall cleanliness of the data. Therefore, we decided to take advantage of the linearity within quasi-linear migration regimes of the graph of correlation with time (see Figure 4.6). We scaled the early-stage correlation with time from each of the seventeen floods to the correlation with time from the nonvegetated stage. (These plots did not contain the point at $C_0 = 1, t = 0$, because this typically plotted off of the line because of effects described in Section 4.3.3.) The result was a straight line, whose slope showed the ratio of the decorrelation rate of the nonvegetated case over the decorrelation rate of the vegetated case. In other words, this slope showed
how many times slower the vegetated runs decorrelated with respect to the nonvegetated run. A plot of decorrelation with time (Figure 6.8) is followed by a graph of slopes (Figure 6.9) on this page and the next.

Figure 6.8. **Individual floods with increasing vegetation density.** F1 through F17 stand for flood number; notice that decorrelation drastically decreases after the first ~6 floods and then remains at approximately the same rate, relative to itself. Each of the curves for the vegetated floods is the curve for the time-step at the beginning of that flood. The curve for the nonvegetated stage is the characteristic exponential decay curve to approximate overall nonvegetated correlation with time.
Figure 6.9. Decorrelation rates with increasing vegetation density: 2002 Experiment (Braided and Vegetated Dynamics) scaled to the nonvegetated case. Note that the slopes of the lines increase as the flood number increases. During each flood, the flume was reseeded at a density of 1 seed per square centimeter, and allowed six days of low flow for vegetation growth (Tal and Paola, 2007).

6.4.2. Analysis

When the slopes of all of these lines are defined, a general trend of lowering channel mobility to a new equilibrium value can be discerned. This new equilibrium value is 0.1569 times the old characteristic channel mobility, which is approximately 6.5 times slower than the original channel mobility in the nonvegetated and braided case. Our
analysis saw only channel migration, not avulsion, so therefore this is a ratio of migration rates in terms of the characteristic channel mobility of the decorrelation analysis. This ratio is significant, and we would like to do further research into how this scales to field measurements. Unfortunately, field measurements are performed by a variety of methods that do not include the decorrelation analysis that we developed. Therefore, we must first compare the methods of analysis, and then compare the results.

**Figure 6.10.** Channel mobility vs. vegetation density (i.e. flood number). An exponential decay curve is fitted to the ratio of mobility rates for each of the floods versus the nonvegetated case. The curve roughly matches the data, which show decreasing migration with increasing vegetation density until a new steady state was achieved at a lower channel mobility. This new mobility is 0.1569 of the old channel mobility, or approximately 6.5 times slower.
List of Terms

\( a \) = The constant in the exponential decay function that determines how much the function is vertically stretched.

\( c \) = The constant in the exponential decay function that determines the location of the horizontal asymptote. When used with random-scaled data, this \( c \) determines the level of permanent confinement on the flow in the system, which is the amount of flume area that the channel does not cross.

\( M \) = The decay constant in the exponential decay function, which determines relative decorrelation rates as well as the curvature of the plot.

\( R^2 \) = Coefficient of determination

\( \frac{d\eta}{dt} \) = Aggradation rate (not including subsidence: just from sediment deposition)

\( M_{HR} \) = Mobility within the high-resolution delta basin experiment (DB03-1)

\( M_{LF} \) = Mobility within the low-Froude-number experiment (DB03-2)

\( M_{Agg,HR} \) = Mobility within the high-resolution delta basin experiment (DB03-1), scaled to aggradation within DB03-1

\( M_{Agg,LF} \) = Mobility within the low-Froude-number experiment (DB03-2), scaled to aggradation within DB03-2

\( \epsilon_o \) = Volumetric sediment density \((1 - \text{porosity})\)

\( Q_s \) = Volumetric sediment flux \((L/s)\); volume given includes grains and pore spaces

\( Q_{s,in} \) = Volumetric sediment flux \((L/s)\); volume given includes grains and pore spaces

\( Q_{s,ou} \) = Volumetric sediment flux \((L/s)\); volume given includes grains and pore spaces

\( A_{\text{delta}} \) = The map-view area across which deposition is occurring; for deltas in which base level is held constant, this area is also held constant

\( M_{Agg} \) = Mobility \((M)\) scaled to the aggradation rate \( (d\eta/dt) \) within its respective experiment.

\( M_{BL\text{const}} \) = Mobility within the base-level constant phase of the cohesive sediment experiment

\( M_{BL\text{rise}} \) = Mobility within the base-level rise constant phase of the cohesive sediment experiment

\( C_{\Phi} \) = The randomness-scaled correlation, or fraction of correlated pixels, scaled to \( \Phi_d \)

\( t \) = experimental runtime; typically taken such that \( t = 0 \) when the baseline time step is being analyzed (and \( C_{\Phi} \) therefore equals 1)

\( \text{FX} \) = Flood number \( X \), where \( X = 0-17 \), in the braided and vegetated dynamics experiment. \( X = 0 \) signifies the nonvegetated state, and \( X = 1-17 \) signifies one of
the floods during the vegetated sequence. Vegetation was added and allowed to
grow in the time in-between each flood.

\[ V = \text{Correlation with time from a specified flood (including F0)} \]
\[ N = \text{Nonvegetated correlation with time} \]
Chapter 7: Applications

7.1 Deltaic Stratigraphy

Deltaic stratigraphy is made up of two major types of fluvial deposits: channel deposits and overbank deposits. The relative abundances and specific characters of these deposits are determined to a great extent by channel migration rate. Therefore, information on the migration rates of channels can shed new light on the development of the fluvial stratigraphic record.

We present a simple method for the comparison between depositional record and channel mobility, which is the same model that was developed by the work of Martin et al., manuscript in preparation, 2007. The only difference is that Martin et al. (manuscript in preparation, 2007) uses timescales, while we use rates.

We create a nondimensional ratio \( R_{\text{strat}} \) between aggradation and scour, in which \( d\eta/dt \) is aggradation rate, \( M \) is the decay constant from the randomness-scaled correlation (with units of 1/time), and \( h \) is a characteristic scour depth:

\[
R_{\text{strat}} = \frac{M \cdot h}{d\eta/dt} \tag{7.1}
\]

We investigate this ratio in order to understand the effects of the different migration rates of DB03-1 and DB03-2 on the stratigraphy they produce, and to compare this with the actual data. The constant \( h \) scales with water flux (Parker et al., in press, 2007) The
same water flux was used in DB03-1 and DB03-2, so for comparison between these two experiments, it is necessary to look only at the ratio of the decay constant and aggradation. This has already been calculated in Chapter 6, and shows that \( R \) for DB03-2 is 4 times \( R \) for DB03-1. This agrees with the observed fact that DB03-2 has a much larger fraction of channel deposits than DB03-1 (Martin et al., manuscript in preparation, 2007).

This value does not scale linearly to the amounts of each deposit that one sees in the deltaic stratigraphy, because channels can rework both channel deposits and sheet-flow deposits. Therefore, making the gross simplifying assumption that channel motion is completely random, the stratigraphic volume of channel deposit should go as an exponential decay with \( R_{strat} \) as the decay constant. However, this is not likely to be very accurate. First, channel motion is not random, as Sheets et al. (2002) and Sheets et al. (in press, 2007) have shown. Channel motion is attracted to different locations on the deltaic surface. In addition, Sheets et al. (2002) show that ephemeral flows are more important for deposition and deeply channelized flows are more important for sediment transport and erosion, and that these deep channels often experience many re-occupations. On top of this, the work of Bryant et. al. (1995) show that there may exist a nonlinear relationship between deposition rate and channel mobility. Therefore, we do not seek to take this argument any further than the following table (Table 7.1) and the accompanying method for calculating the exponential decay curves. We take an arbitrary scour depth, \( h \), of 1 cm. Using this value of \( h \), \( R_{strat} \) for DB03-1 and DB03-2 are shown in Table 7.1.
Table 7.1. $R_{strat}$ and volume percentage of channel deposit for DB03-1 and DB03-2.

<table>
<thead>
<tr>
<th></th>
<th>$R_{strat}$ (with $h$ in meters)</th>
<th>$R_{strat}$ (with $h$ in mm)</th>
<th>Approximate Volume Percentage of Channel Deposit</th>
</tr>
</thead>
<tbody>
<tr>
<td>High-Res (DB03-1)</td>
<td>0.002629</td>
<td>2.629</td>
<td>~60%</td>
</tr>
<tr>
<td>Low-FR (DB03-2)</td>
<td>0.01230</td>
<td>12.30</td>
<td>~95%</td>
</tr>
</tbody>
</table>

Approximate values for volumetric percentages of channel deposit are as given in the work of Martin et al. (manuscript in preparation, 2007). These approximations for stratigraphic volume are vague estimates, and should not be taken as hard data.

Note that DB03-2 had more wetted area than DB03-1. We are assuming (not without some basis) that a balance exists between wetted area and scour depth, such that as wetted area increases, scour depth decreases to balance channel cross-sectional area. Therefore, we believe that the channels in both experiments will have the same net volume.

Although our values are not concrete enough to perform this calculation ourselves, we hypothesize a method for matching mobility and aggradation to the proportions of channel deposit and sheet-flow deposit that we see in the stratigraphy. First, we define the deposit volumetric ratio $V_R$ to equal volume of channel deposit $V_c$ divided by total deposit volume $V_{tot}$.

$$V_R = \frac{V_c}{V_{tot}}$$

(7.2)

Then, we set up the exponential decay:

$$(1 - V_R) = e^{-\gamma R_s}$$

(7.3)

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In this equation, $q$ is the decay constant. This is the final form of our proposed simple equation to compare aggradation-scaled mobility and proportions of channel and sheet-flow deposit.

### 7.2. Environmental Management

#### 7.2.1 Farmland Soil Loss and the Resulting Transport of Soil and Nutrients

Land loss due to erosion in agricultural areas is a major problem for both the farmer and the natural resources manager. Farm erosion removes good soil from agricultural land and sends this silt into the watershed along with its nutrients ($K, NO_3, PO_4$), which are often from man-made fertilizers and can cause algal blooms. These problems are often the result of poor riparian area management during agricultural use (Peppler and Fitzpatrick, 2005).

As of current measurements, agricultural areas are experiencing increased rates of channel migration (Peppler and Fitzpatrick, 2005; Knox, 2001). Planting crops in place of native vegetation increases floodplain erodibility by 80-150% (Micheli et al., 2004), and cattle trampling stream beds increases erosion (Peppler and Fitzpatrick, 2005). Because of this increased erosion, streams are carrying additional sediment, and this sediment is depositing in lakes along the course of the river. This problem is multiplied because most rains fall in spring, when no crops are planted on fields, and the channels are able to often
erode bare soil, which is neither reinforced by roots from erosion nor protected by leaves from precipitation. This often results in gully formation, which robs farmers of precious farmland, and sends the soil as sediment in streams and lakes that exist in an equilibrium without it. In addition, erosion produces a chemical problem. When rivers in cropland erode their banks, they entrain anthropogenic fertilizers, increasing the nitrate or phosphate content of the water and causing eutrophication (Lung and Larson, 1995). This causes the growth of algal mats, which use large amounts of dissolved oxygen to respire at night, thereby killing the native aquatic life that depend on oxygen in the water (Lung and Larson, 1995).

Lake Pepin, a natural lake on the Mississippi River on the Minnesota-Wisconsin border, is filling in fast with sediment and anthropogenic phosphorus due to farming along the Minnesota River valley. Kelley et. al. (2000) shows that sediment influx into Lake Pepin from the heavily agriculturalized Minnesota River has increased twelve-fold since the beginnings of farming along its banks. This sediment is rapidly filling the lake, and bringing nutrients with it, such as phosphates, that cause algal blooms and eutrophication. James et al. (1995) show that Lake Pepin’s sediments contain remarkably high phosphorus concentrations.

New farming practices that include better riparian management need to be initiated in order to curtail these problems of soil erosion, soil transport and deposition, and nutrient transport and deposition. Understanding the mobilities of channels with and without riparian vegetation is essential for a better understanding of how to manage streams on
farmlands in a way that is sustainable to both the farmer and the natural resource manager. For now, the message seems clear that native riparian vegetation is good for erosion protection. Additional work in the laboratory and in the field should prove helpful in their ability to constrain erosion rates and balance sediment budgets.

### 7.2.2 Riparian Ecosystem Management

Maintaining channel mobility is essential for the health of riparian habitats (e.g., Ollero, 2007). Many studies show that lower discharges, which are becoming preeminent in rivers around the globe due to flood control projects, change migration rates, morphology, and ecology of riparian zones (Tal et al., 2004; Ollero et al., 2004; Ollero, 2007; Micheli et al., 2004). This change is often detrimental to regional ecology and endangered-species management. For example, Scott et al. (1996) that floods are required for cottonwood establishment. In other studies, destruction of riparian vegetation leads to faster channel migration, which can cause problems for local ecology as well as for human development in the riparian zone (e.g., Knox, 2001; Micheli et al., 2004). This need to strike a balance between what is necessary for human society and what is necessary to keep the ecosystem in balance makes land-use planning and sustainable development around river systems especially difficult.

The recent state of understanding of the controls of riparian vegetation channel mobility has not been uniform. Micheli et al. (2004) state that traditional strategies for flood control generally include the removal of riparian vegetation, while Simon and Collison
(2001) report that riparian vegetation is used by stream managers to maintain bank stability. The variation in methods, as shown in these two different publications, demonstrates that there has been no good consensus on how to use riparian vegetation to stabilize rivers in the recent past. However, the general consensus today is that riparian vegetation slows channel migration and helps to stabilize a riparian zone.

Thanks to increased research in quantifying channel migration (e.g., Micheli et al., 2004; this work), bulk morphological changes (e.g., Gran and Paola, 2001; Tal et al., 2004; Tal and Paola, 2006) and bank stability (e.g., Simon and Collison, 2001), we now understand that riparian vegetation increases bank strength through rain interception and root cohesion, and that this increased strength outweighs the loss of stability caused by the weight of the plants and the changes in morphology induced by large woody debris. Increased work in quantifying channel migration will continue to aid in our understanding of riverine processes and the rates at which they occur.

7.3. Alluvial-Fan Hazards

Alluvial fans are good locations for humans to live because they provide large, gently dipping expanses of land within places that are otherwise dominated by rugged mountains. They are a bad place for humans to live because the sedimentary processes that shaped these fans are occurring and reshaping them on a human time scale. As more people move into the American Southwest, it becomes increasingly important to
understand how these fans operate in order to understand how to keep these people safe, or to dissuade them from living on the fans.

The main channel (or channels) in a channelized alluvial fan act as conduits for sediment. The depositional record and a simple application of the depth-slope product tell us that the sediment size can be anything from sand to boulders. By using our experiments with alluvial fan deltas, it may be possible to gain a better understanding of rates of channel change on channelized alluvial fans that allow for safer management of alluvial fan communities.

We investigate one extremely noteworthy disaster, which occurred in Vargas state, Venezuela, in December of 1999. Large floods caused debris flows which roared through highly urbanized alluvial fans along the Caribbean coast, just north of Caracas. These debris flows cut channels across the fans that claimed an estimated 19,000 lives. Because the land is so steep here, the alluvial fans are the only places where it is possible to build. Therefore, it is important to understand where the channels are, and how rapidly they are likely to change location, in order to properly assess risk and emplace policy in order to avoid future loss of life. Because of our experiments on alluvial fan deltas, which are very similar to situations such as this, we may be able to use our results for channel mobility with respect to sediment flux to understand channel mobility rates in hazard zones like these fans.
Figure 7.1. **Looking North over Caraballeda, Venezuela, an alluvial fan city.** (Photo Courtesy of Matthew Larsen, as published by Larsen et al., 2001.)

Figure 7.2. **Looking South over Caraballeda, Venezuela.** Note the well-defined channels that sweep through the city, demolishing some parts and leaving other parts almost unharmed. (Photo Courtesy of Matthew Larsen, as published by Larsen et al., 2001.)
Figure 7.3. **Mud-line.** Buildings display a mud-line from the debris flow in the shape of the flood deposits. The street going uphill served as a main channel for the debris flow. (Photo Courtesy of Matthew Larsen, as published by Larsen et al., 2001.)

Figure 7.4. **Destroyed apartment building.** This apartment building lost its side and first two stories to boulders transported in the debris flow through Caraballeda, Venezuela. Notice the channel to the right of the image, and the drastically increased amount of damage to the part of the building in the channel’s path. (Photo Courtesy of Matthew Larsen, as published by Larsen et al., 2001.)
7.4. Hazards of Aggradation and Avulsion

Channel migration and avulsion is hotly debated in the field of science and public policy. In the United States, management of the lower Mississippi River is highly controversial. These debates often center on the avulsion into the Achafalaya River that the Army Corps of Engineers stalled with the construction of the Old River Control Structure and the progradation of the artificial bird’s-foot delta while the wetlands remain sediment-starved. The recent catastrophe of Hurricane Katrina underscores problems with river and sediment management in the Mississippi delta. However, problems due to aggradation and avulsion have also struck other parts of the country, such as the town of Niobrara, Nebraska.

In 1957, the Gavins Point Dam was installed on the Missouri River, gradually raising base level on the Niobrara River due to deposition induced by the presence of the Lewis and Clark reservoir. This aggradation and avulsion destabilized the channel system, covered a state park in sediment, and resulted in the evacuation and relocation of the town of Niobrara (Bristow et al., 1999; Jerolmack, 2007).

This example, which illustrates enough of a disaster in its own right, also has the interestingly applicable quality of being related to base-level rise. Global sea level is expected to rise, and this effect will be felt in deltaic channel systems, which will tend to backfill as their backwater readjusts to the new conditions. This backfilling will increase superelevations and the potential for avulsion, especially on shallow-sloped, low-lying deltas.
In coastal cities like New Orleans, USA, subsidence and base-level rise combine to create a condition of rapid relative base-level rise. If these trends of subsidence and sea-level rise continue, rise of relative base level, combined with the shallow slope of the Louisiana bayou, will cause much of the Mississippi delta to become inundated. The bed of the Mississippi River will aggrade to come to equilibrium with the new base level. Back-filling increases superelevation, producing ever-increasing costs in levy construction and maintenance, and ever-increasing risks of catastrophic floods. Our results show these predictions to be correct, and that a rapid increase in sea level should cause channels in coastal regions to become mobile at 1.5 to 3 times their usual rate.
List of Terms

$R_{strat} =$ A characteristic nondimensional stratigraphic ratio between channel deposit and sheet-flow deposit

$M =$ The decay constant in the exponential decay of randomness-scaled correlation, with $a$ and $c$ held constant, that gives the characteristic channel mobility

$h =$ Characteristic channel scour depth

$\frac{d\eta}{dt} =$ Aggradation rate (not including subsidence: just from sediment deposition)

$V_c =$ The volume of stratigraphic deposit that is channel deposit

$V_{tot} =$ The total volume of deposited stratigraphy

$V_R =$ The ratio of $V_c / V_{tot}$, or the volumetric proportion of the stratigraphy that is channel deposit

$q =$ The characteristic decay constant for the proposed exponential decay relation of $V_R$ and $R_{strat}$. 
Chapter 8: Discussion and Conclusions

8.1. Discussion

This thesis contains two parts. The first is a description of the method we used, and our attempts to understand how pixel decorrelation relates to channel migration. The second is an application of pixel decorrelation to understanding experimental alluvial environments, with a few first-order field-scale extrapolations. In this section, we review the work in this thesis and look at how it relates to our understanding of channel mobility in general through a review of some of the preexisting literature.

8.1.1. Methods and Theory

We showed three ways in which decorrelation is related to channel migration. First, the overall form of a curve of correlation (the opposite of decorrelation, scaled to either randomness or flume surface area) with time is that of an exponential decay curve that is characteristic to the general channel mobility of the particular system that we are studying. Second, features that deviate from the exponential decay curves and that exist without regard to the chosen baseline time step show different regimes of channel location. Third, looking at two subsequent images quantifies instantaneous decorrelation in magnitude and in time, which can be used to find larger-than-normal fluvial events.

We almost exclusively used the first of these ways, by selecting baseline images and measuring correlation with time from that baseline image. From these, we found characteristic channel mobilities, which to physical parameters within the fluvial systems
in order to better understand the interactions between these parameters and channel mobility.

We would like to state that we have not definitively proven that this decorrelation follows an exponential decay curve. The data suggest something that looks like exponential decay, but there are many other curves that can roughly match the same shape, especially including those of rational functions. For now, we have shown some theory behind why we believe the exponential decay curve to be a correct portrayal and what the various constants in the equation show in terms of fluvial system dynamics. In addition, for the geological applications, exponential decay is a curve that fits the data and is workable. Therefore, it is sufficient for the empirical relationships developed in part 2 (Chapters 5-7). It is our hope that new and better insights into fluvial processes will allow us to understand whether or not our hypothesis of exponential decay is correct. If it is correct, we hope that future research will allow us to explore the deeper physical meaning behind the curve characteristic to a particular channel system.

8.1.2. Results and Applications to Fluvial Systems

We studied the effects of deposition due to incoming sediment flux, deposition and erosion due to base-level changes, and bank cohesion on channel mobility. In each of these situations, we calculated ratios of the channel mobilities with respect to different extrinsic variables.
Our idea that channel mobility should be related to deposition is supported by current research. Work done by Parker et al. (in press, 2007) shows that channel cross-sectional area should scale with $Q_w$, and work done by Mohrig et al. (2000) and Jerolmack and Mohrig (2007) show that channel filling should scale with aggradation. Therefore, looking at a ratio of these two values gives an idea of the channel mobility with respect to volumetric water and sediment fluxes. Our work agreed with this argument to an extent: the channels in DB03-1 were mobile at 51 times the rate of the channels in DB03-2. However, our work shows that channel mobility can involve more factors than a simple filling argument can satisfy. In particular, we believe that the existence of flow-depth-high ripples and shallower, broader flows due to a lower slope during DB03-2 caused its channels to migrate more rapidly with respect to its aggradation rate than the channels in DB03-1 did with respect to its aggradation rate.

Field examples such as the Niobrara River in Nebraska provide evidence that rising base level causes a knickpoint to migrate up the long profile of a river (Bristow et al., 1999). In the case of an alluvial river, this knickpoint migrates by deposition as the zone of still water moves upstream, and causes the channel’s backwater to move upstream with it. This aggradation causes superelevation, and as per the work of Bryant (1995), Heller and Paola (1996), and Mohrig et al. (2000), the river becomes avulsive. This is a significant concern in a place like the Niobrara River, where a state park and the town of Niobrara were relocated to higher ground in response to flooding and avulsion due to aggradation from downstream base-level rise behind the Gavin’s Point Dam (Bristow et al., 1999). What is even more potentially significant is the effect of aggradation on the Mississippi
River delta as global warming causes global sea-level rise. Our results showed a change in channel mobility of 1.5 times faster during base-level rise than during times when sea level is constant, and an increase of channel mobility by a factor of 3 between rapid sea-level fall and sea-level rise. Depending on the state of the oceans with respect to river-channel time scales, and especially considering effects on rivers and sea level from the last glacial maximum, our models would seem to suggest an increase in the mobility of the Mississippi River by some magnitude of the order that we have seen in our experiments. Whether this is actually the case or not is unknown, inasmuch as the Mississippi delta is a complex system, and we have not rigorously scaled the experiments. However, at the very least, we believe that coastal rivers will tend to be more mobile with sea-level rise.

Understanding channel mobility in vegetation as opposed to in noncohesive sediment is important in its applications to geological studies of river system-behavior and to stream management and restoration. Gran and Paola (2001), Tal et al. (2004), and Tal and Paola (2007) have assembled a wealth of experimental data on the morphological changes that occur when a sandy flume is planted with alfalfa (Medicago sativa), including narrowing and deepening of the channels and an overall change from a braided plan form to a more sinuous and/or anastomosed plan form. Our findings indicate that our characteristic value for channel migration decreases by a factor of 6.5 from the fully nonvegetated to the fully vegetated stages of the experiment.
8.2. Conclusions

- We show that spatial change can be used as an effective metric by which to measure channel network migration and avulsion, and to characterize channel change with respect to extrinsic parameters. This method is especially suited to braided environments because it is not dependent on rigorously defining channels. It works particularly well with experimental data because experimental data sets are dense and because most of the extrinsic variables on the system are defined.

- The mostly noncohesive delta, XES02 (Kim et al., 2006; Kim, 2007), showed no noticeable change in channel migration between base-level fall and rise that was slow with respect to channel mobility. During base-level fall and rise that was fast with respect to channel mobility, channel mobility during base-level fall was three times channel mobility during base-level rise.

- The 2007 cohesive sediment experiment (Hoyal and Sheets, manuscript in preparation, 2007), utilizing the new cohesive sediment mixture, showed a 1.5x increase in channel mobility during base-level rise that matched sediment input to keep a constant subaerial delta surface, as opposed to mobility while base-level was held constant.

- In the 2003 run of the braided and vegetated dynamics experiment (Tal and Paola, 2007), the channels migrated 6.5x more slowly during the vegetated steady state than during the nonvegetated steady state. This is illustrative of how riparian vegetation controls river migration in field settings, as well as of the slower migration rate of sinuous rivers as opposed to braided rivers.
• Channel mobility increases with increasing sediment input and aggradation. This is seen in the fact that the channels in DB03-1 (Sheets et al., in press, 2007) were more mobile than the channels in DB03-2 (Martin et al., manuscript in preparation, 2007). This helps to verify the work of Bryant et al. (1995), Heller and Paola (1996), and Mohrig et al. (2000). However, channel mobility does not increase linearly with increasing sediment flux and aggradation. Shallow surface slope in DB03-2 caused sheet flow, and ripples in DB03-2 that scaled with flow depth acted to divert the flow (Martin et al., manuscript in preparation, 2007). Therefore, when scaled to their respective aggradation rates, DB03-2 was actually mobile at about four times the rate of DB03-1. These results are consistent with the fact that we see more channel deposit in DB03-2 than in DB03-1, and this deposit ratio can viewed in terms of a nondimensional ratio of channel mobility and scour depth against aggradation rate.

• DB03-1 shows significant periodicity in its occupation of different zones on the delta surface.

• Reductionist analyses such as this can shed light on the interactions between physical parameters in the environment and sedimentary processes. This information can then be taken back into the field, either in geology and geography, or into environmental and natural hazard management.
8.3. Future Work

We would like to pursue future work on this project in order to examine other features visible in the decorrelation analysis, look more deeply into the physical reasons for channel mobility, apply our method to field-scale examples, compare our results to field studies, and investigate possible applications of this method to geology and other scientific and environmental fields. One goal is to do a full investigation of instantaneous decorrelation. We would also like to do a full investigation into the constant-in-time features in the plots, in order to find preferential channel regimes over time. (We have started some of the theoretical portion of this work.) We are particularly interested in depositional pattern on delta surfaces and what attractors or repulsors are controlling the periodicities we see in fluvial activity. Another goal is to use decorrelation analysis to further our understanding of channel mobility with respect to varying depositional, geographical, and fluid mechanical conditions. A fourth goal is to compare this work to field studies. This will be done in two ways. First, we will apply our method of decorrelation analysis on a time series of aerial and/or satellite photos of a channel system that is moving at measurable speeds and for which as many of the major parameters (sediment flux, slope, water flux, width, depth, grain size distribution, etc.) are known. Second, we will compare our results with channel mobility to other methods for quantifying channel mobility (such as meander-bend migration and centerline deformation) to gain an appreciation of the relationship between these and decorrelation time scales. (This relationship could also be attained in an experimental setting by using centerline deformation and meander-bend migration to characterize the experiments that we have already analyzed.) Our fifth goal is to more deeply investigate possible
applications of this work, in (not limited to) alluvial fan hazards, deltaic stratigraphy, continental alluvial stratigraphy, paleoenvironmental reconstruction, geoarchaeology, and environmental management and planning in watershed and riparian areas.

8.3.1. Effects of the Environment on Channel Mobility

One of our major goals is to understand channel mobility as a recorder of the physics and environmental conditions of the past and present. Channel mobility is affected by sediment flux, water flux, bank stability (cohesiveness), channel geometry, and base level, to name a few factors. Our experimental work provides an avenue to develop datasets that pick apart the different extrinsic variables. This allows us to isolate their effects on the channel system and approach this problem as reductionists. Eventually, and with enough success, it may be possible to use channel-mobility data as an additional tool to measure paleoenvironmental conditions.

8.3.2. Present-Day Environmental Concerns

Our first goal in addressing environmental concerns is to find groups who care about channel migration in an environmental context, and to understand what these groups would like from the scientist's perspective. After this, we will proceed to create an applicable version of our research. One possible idea for an application of the scaling relationships that we are developing to measure channel mobility involves measuring bank cohesiveness, sediment flux, normal-flow water flux, and frequency of overbank flows. The researcher will then use our relationships to extrapolate to channel mobility.
Of course, this idea is still just a concept, and the results of our research as well as the needs of the environmental groups will alter our final product.

8.3.3. Additional, Specific Goals

In addition to the above-stated goals, we have five other more specific outstanding questions. In particular: (1) “What is the effect on channel mobility of vegetation inside channels?”, (2) “What is the interaction between sediment flux, water flux, Froude number, deltaic slope, base level and channel mobility?”, (3) “Can we define stratigraphy in terms of random channel motion on a deltaic surface, along with aggradation rate, scour area, and channel mobility?”, (4) “How does sediment flux affect channel mobility in systems in which sediment input equals sediment output?” (In other words, what is the relationship between non-aggradational channel mobility from sediment flux?), and (5) “How do channel mobility and rates of base-level change interact?” Work has started insofar as addressing the second, third, fourth, and fifth questions, but our data points are few and the potential nonlinearities are many, so we believe that drawing any conclusions would be too premature at the present time.
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