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Exploring Predictive Relationships of Fluvial Morphology: Using Shuttle Radar Topography Mission Data

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Exploring Predictive Relationships of Fluvial Morphology:

Using Shuttle Radar Topography Mission Data

by

Mark Thomas Hannon

B.A., University of Colorado, Boulder, CO 2007

A thesis submitted to the Faculty of the Graduate School of the University of Colorado in partial fulfillment of the requirement for the degree of

Master of Science

Department of Geology

2011
This thesis entitled:

Exploring Predictive Relationships of Fluvial Morphology:

Using Shuttle Radar Topography Mission Data

has been approved for the Department of Geology

_________________________________________
James P.M. Syvitski

_________________________________________
Albert J. Kettner

The final copy of this thesis has been examined by the signatories, and we find that both the content and the form meet acceptable presentation standards of scholarly work in the above mentioned discipline.
To identify general large-scale patterns (slope, slope change, sinuosity) along a river’s course the worldwide SRTM 3 arc-second DEM satellite derived data was analyzed. Longitudinal profiles were calculated for sixteen rivers. This analysis uses auxiliary data sets to develop an understanding of the external and internal influences that are pressed upon and inherent within the lower 100 meters of the river systems. Contradictory to previous findings, the sixteen rivers studied here show that slope and sinuosity are not strongly correlated at the reach scale. The total river's longitudinal profile up to 100 meters, provides an average slope and sinuosity throughout the entire system and increases the correlation between slope and sinuosity (~0.56). Comparing the entire river's longitudinal profiles also illustrates a threshold of planform sinuosity (>1.6) in which meandering rivers are found. Using this threshold, the Indus, Mississippi, and Fly Rivers are further examined to understand lateral migration rates, the link between meandering rivers and the production of oxbow lakes throughout their floodplain. The slope of three rivers was examined for external controls by overlaying geological data of bedrock type and fault locations. Neotectonics appears to impact the slope and/or sinuosity of the Mississippi, Niger, and Magdalena rivers. Results indicate growth faulting found in the mud-dominated systems of the Mississippi and Niger influences sinuosity. The resulting sinuosity is greatest in regions where these rivers are bound by growth faults. The Magdalena has several regions where the river intersects strike-slip faults,
resulting in increased slopes with the more parallel the encounter. River longitudinal profiles can also reveal areas of bedload erosion and deposition. Zones of erosion (sources) and deposition (sinks), and knowing how to locate them, are of great interest to a variety of geoscientists. These predictive relationships will provide future assistance to the field of fluvial morphology.
Acknowledgments

I would like to first and foremost thank my committee members. Professor Syvitski provided the funding, overarching goals, and expertise to complete this research. Dr. Kettner was a guide through the endless web that is GIS and programming. Dr. Overeem was always there for a swift kick in the right direction. The Delta Force delivered many answers to the questions I had throughout my studies. Thanks to Conoco for the funding support and ideas. My family and friends provided much needed laughs and support throughout my studies. Finally, I want to thank Katie for her support and patience throughout my research. Thanks to you all.

Dedicated to Charles J. Fox.

-For you Jimmy, wherever you are.
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Chapter 1 Introduction

Satellite data obtained by the United States is freely available, and allows for analysis and comparison of the Earth by remote sensing without disturbing the environment. The Shuttle Radar Topography Mission (SRTM) produced an almost worldwide digital elevation model (DEM) allowing for the break away from traditional field-based techniques used to examine rivers on a case-by-case basis. The DEM is noted for its global consistency in terms of data quality and the short period of data collection.

This study focuses on lowland fluvial floodplain systems. Examining floodplain architecture for the 16 major river systems with an objective to develop predictive relationships for longitudinal profiles, as well as planview characteristics of river floodplains. With over half a billion people living on deltas and the rivers that create, shape, and dominate these regions, there is a need to develop predictive relationships on the processes that control floodplain architecture (Syvitski et al., 2009).

The longitudinal profiles of 16 rivers throughout the world are obtained, resolving the local convexities and concavities of the rivers downstream slope (Figure 1.1). These local anomalies of downstream slope provide insight into the physical nature surrounding the cause of the fluctuations in slope (i.e. lithology and faults). Due to difficulty in determining what separates coastal lowlands from a delta, the study examines rivers up to their 100-meter contour (Syvitski, 2008). The 100-meter upper limit also allows for an examination of the strictly fluvial connection to the lower tidal influenced deltas. Many
studies have been performed, attempting to pinpoint the causes of changes in slope using case-by-case studies (Knighton, 1998; Snow and Slingerland, 1987; Gregory and Schumm, 1987), but have not been done so with a global scale. This study sheds light on the group of selected rivers, providing their slope and sinuosity, proving the ability to examine all rivers at a near global scale. The research allows for a correlation analysis of slope and sinuosity encompassing a range of fluvial environments. Channel slope and channel sinuosity are examined at scales of sub-reach, reach, and entire river profile.

Different lithological regions give rise to profiles of different concavity, implying differing rates of yielding to their final equilibrium state (Brush, 1961). Faults introducing fractures and can increase the erosion rate with the transport power of a river moving through (Waclawik et al., 2008). Knighton (1998) found that slope at any point is related to bed material size, discharge and sediment load. If all of the variables (discharge, sediment load, lithology, and faults) indeed affect local slope and concavity, then there should be a visible physical parameter controlling the abnormal change in slope downstream, making them apparent in the rivers’ longitudinal profiles. Encompassing different effects of climate zones, lithologies, and tectonics, this group of rivers is related in a search for a comparative link between sinuosity and slope.

The sixth chapter discusses the use of SRTM-calculated slope applied in a bedload equation. Bedload transport is responsible for in-stream habitat complexity and preservation, as well as deltaic accretion, and the ability of a river to recover from natural or anthropogenic disturbances, including floods and upstream impoundments (Gomez and Church, 1989; Knighton, 1998). Obtaining freely available discharge data for the
Mississippi and Magdalena Rivers, from the USGS and IDEAM respectively, combined with the acquired slope from SRTM bedload is calculated at the reach scale.

In Figure 1.2 the SRTM DEM of the lower Mississippi River’s basin is shown, both with a typical color schema and a cyclic color schema. To illustrate the resolution of the SRTM DEM Figure 1.3 shows a zoom in of Figure 1.2, allowing for easy identification of oxbow lakes and anthropogenic features (i.e. agricultural plots). Finally, Figure 1.4 represents a cross-sectional profile taken at the apex of the Yellow River. Note the high-hardened levees towering 10+ meters above the surrounding floodplain show up in the 90-meter resolution SRTM DEM.

Figure 1.2 Lower Mississippi River Alluvial Basin. A) Typical color schema from Blue (sea level) to White (mountain peaks). B) Cyclic color schema with colors repeating every 10 meters until 100 meters in elevation is reached (black thereafter).
Figure 1.3 Zoom in from Figure 1.2 of the Lower Mississippi River Alluvial Basin.  A) Typical color schema (Fig. 1.2 A).  B) Cyclic color schema (Fig. 1.2 B).  C) Landsat Imagery, finer resolution than the SRTM. Note how the cyclic color schema allows for easy identification of oxbow lakes and agricultural plots (Best of both worlds.)
This study addresses causation of local anomalies of slope in a river’s longitudinal profile. How does floodplain longitudinal profiles differ from longitudinal profiles encompassing an entire river drainage basin? Does river slope explain the variance of a river’s planform sinuosity? At what scale can we compare river systems? Can a threshold sinuosity be established for a meandering system? Does the high resolution SRTM DEM show an expression of a fault or bedrock transition in a river’s longitudinal profile? Can bedload fluxes be determined from the slope of the SRTM DEM and do they match well with fluxes determined from literature? Finally, since the SRTM is a snapshot of 2001, can we use historical maps to determine lateral migration rates for the Indus River system? Can we define predictive relationships for oxbow lake occurrence in the floodplain?

Figure 1.4 A cross-sectional profile taken at the apex of the Yellow River. Note the high-hardened levees towering 10+ meters above the surrounding floodplain.
Chapter 2 Data and Methods

In an attempt to understand physical properties of floodplains and deltas, this study utilizes high-resolution satellite data to characterize selected fluvial systems. The research focuses on the floodplain and delta regions of 16 major fluvio-deltaic systems. The rivers are selected to capture a wide variety of fluvial-deltaic systems (Table 2.1). Wave, tide, and fluvial dominated delta systems are represented. Basins encompassing different size and varying degree of river’s sinuosity are selected for the study to show that comparing all rivers is feasible with the methods utilized herein. The study examines the longitudinal profile of the river through the water surface elevation (WSE) up to the first 100 meters using the SRTM DEM to resolve local concavities and convexities through the creation of longitudinal profiles of 16 fluvial systems. Due to the small width of several rivers, and anthropogenic influences (e.g. dams) on several rivers not all longitudinal profiles are mapped to their first 100 meters of WSE as planned.

Examining the selected fluvial systems provides for an update on the early version of the planform classification of Leopold and Wolman (1957), by dividing rivers into three categories: braided, meandering and straight. Schumm and Khan (1972) found that there are clear break points in the three types of river types: straight, meandering, and braided (Figure 2.1). This classification scheme is dependent upon the calculated sinuosity of each river system. Sinuosity and slope are calculated at varying scales along the examined reach, up to 100 meters. The smallest reach, termed sub-reach, is measured
based on the vertical resolution of the SRTM data. The data is broken on 1-meter vertical bins, so the sub-reach is the length of a 1-meter bin on the river (this could be several tens

Table 2.1 Fluvial system and floodplain architectural parameters.

<table>
<thead>
<tr>
<th>River, Country (Lat/Lon)</th>
<th>Drainage Area (km²)</th>
<th>Avrg. Discharge (m³/s) Peak/ Monthly</th>
<th>Architectural description 5-meander Sinuosity Range</th>
<th>Mean Slope (cm/km) and Sinuosity</th>
<th>Climate zone Mean Annual Precip (mm)</th>
<th>Lakes Present</th>
</tr>
</thead>
<tbody>
<tr>
<td>Columbia, US W 124° 00' N 46° 15'</td>
<td>671,973</td>
<td>6,800 17,000</td>
<td>Narrow channel, structurally controlled channel meanders. Numerous dams and tidal dominated estuary 1.1-1.3</td>
<td>18 ---1.2</td>
<td>Temperate 180-450</td>
<td>No</td>
</tr>
<tr>
<td>Eel, US W 124° 15' N 40° 38'</td>
<td>14,208</td>
<td>235 750</td>
<td>Narrow channel, structurally controlled channel meanders.1.2-1.4</td>
<td>68---1.3</td>
<td>Temperate 1,500</td>
<td>No</td>
</tr>
<tr>
<td>Barito, Borneo E 114° 30' S 3° 19'</td>
<td>74,155</td>
<td>350 450</td>
<td>Very wide floodplain. Meandering channels. 1.4-2.2</td>
<td>5 --- 2.1</td>
<td>Tropical 2,500-4,500</td>
<td>Numerous oxbow lakes</td>
</tr>
<tr>
<td>Kikori, Papua New Guinea E 144° 19' S 7° 39'</td>
<td>21,504</td>
<td>1,673 -</td>
<td>Narrow channel, structurally controlled channel meanders 1.2-2.6</td>
<td>36 ---1.3</td>
<td>Tropical 200-11,000</td>
<td>No</td>
</tr>
<tr>
<td>Fly, Papua New Guinea E 143° 30' S 8° 30'</td>
<td>61,413</td>
<td>6,500 7,200</td>
<td>High sinuosity 2-3, except in tidal river. Many abandoned channels with large-scrol bar topography</td>
<td>4 ---2.3</td>
<td>Tropical 200-11,000</td>
<td>Very abundant oxbow lakes, tie-channels, wetlands.</td>
</tr>
<tr>
<td>Brahmaputra, Bangladesh E 90° 47' N 22° 43'</td>
<td>522,476</td>
<td>- 38,000</td>
<td>Brahmaputra as Braided over the largest scale. Many crevasse splays. Many tidal channels in delta 1.3-1.6 (1.1-1.2 Braidplain)</td>
<td>9 ---1.5</td>
<td>Monsoon 2,000-5,000</td>
<td>Yes</td>
</tr>
<tr>
<td>Kapuas, Borneo E 109° 32' S 0° 15'</td>
<td>80,368</td>
<td>2,384 5,668</td>
<td>Very wide floodplain. Meandering channels 1.5-2.5</td>
<td>4 ---2.0</td>
<td>Tropical 2,500-4,500</td>
<td>Very abundant oxbow lakes</td>
</tr>
<tr>
<td>Indus, Pakistan E 67° 44' N 24° 14'</td>
<td>1,143,101</td>
<td>3,000 30,000</td>
<td>Large meandering, sinuosity 1.2-2.0 sandy-bed, active migration.</td>
<td>9 ---1.6</td>
<td>Arid-monsoon 396</td>
<td>Few oxbow lakes, mostly on the delta.</td>
</tr>
<tr>
<td>Guaviare, Columbia W 67° 57' S 3° 58'</td>
<td>107,365</td>
<td>5,500 -</td>
<td>Very wide floodplain. Meandering channels 1.5-2.8</td>
<td>1 ---2.1</td>
<td>Monsoon 1,500-4,000</td>
<td>Numerous oxbow lakes</td>
</tr>
<tr>
<td>Mahanadi, India E 86° 40' N 20° 19'</td>
<td>141,207</td>
<td>2,112 6,900</td>
<td>Hardly any floodplain, braided with 2km wide floodplain, relatively sandy, main channel belt sinuosity 1-1.3</td>
<td>28 ---1.2</td>
<td>Monsoon 1360</td>
<td>Few, on the delta</td>
</tr>
<tr>
<td>Volga, Russia E 48° 30' N 46° 10'</td>
<td>1,476,411</td>
<td>7,600 24,022</td>
<td>Highly meandering anastomosing river system, highly vegetated. Large point bar features 1.1-1.3</td>
<td>3 ---1.4</td>
<td>Temperate 626</td>
<td>Many Compaction Lakes</td>
</tr>
<tr>
<td>Magdalena, Columbia W 74° 48' N 10° 58'</td>
<td>251,743</td>
<td>7,530 10,000</td>
<td>Highly meandering, sinuosity of 1.1-1.4, relatively straight channel in steeper parts of profile. Anastomosing in reaches where channel encounters strike-slip faults</td>
<td>13 ---1.4</td>
<td>Tropical 2,050</td>
<td>Many lakes in tectonic depression.</td>
</tr>
<tr>
<td>Location</td>
<td>Area (km²)</td>
<td>Population (thousands)</td>
<td>Description</td>
<td>Sinuosity (1.3-1.8)</td>
<td>Climate</td>
<td>Observations</td>
</tr>
<tr>
<td>----------------</td>
<td>------------</td>
<td>------------------------</td>
<td>-------------------------------------------------------------------------------</td>
<td>----------------------</td>
<td>------------------</td>
<td>---------------------------------------------------</td>
</tr>
<tr>
<td>Mississippi, US</td>
<td>3,202,959</td>
<td>15,452 30,482</td>
<td>Prominent meandering (up to local stretches with sinuosity of 2.5 in growth fault bounded sub-reaches) Large scroll bar geometry 1.3-1.8</td>
<td>6 ---1.7</td>
<td>Temperate, subtropical</td>
<td>1,600 Very abundant oxbow lakes, tie-channels, wetlands.</td>
</tr>
<tr>
<td>Kura</td>
<td>218,906</td>
<td>550 2,250</td>
<td>Meandering, large floodplain 1.7-2.3</td>
<td>5 ---2.0</td>
<td>Temperate, subtropical</td>
<td>200-500 Few oxbow lakes.</td>
</tr>
<tr>
<td>Niger, Nigeria</td>
<td>2,240,019</td>
<td>6,130 12,000</td>
<td>Channel sinuosity varies 1-1.2, stretches of 1.7 in growth fault bounded sub-reaches near delta apex.</td>
<td>9 ---1.5</td>
<td>Tropical</td>
<td>690-2,850 Yes</td>
</tr>
<tr>
<td>Yellow, China</td>
<td>983,627</td>
<td>1,480 28,603</td>
<td>Abandoned, super-elevated channel belts Many crevasse splays. 1.1-1.5</td>
<td>12 ---1.3</td>
<td>Several climate zones 450</td>
<td>Few e.g. Hongze Lake, Nanyang Lake</td>
</tr>
</tbody>
</table>

Sources* (Syvitski et al., 2005; Meade et al., 1990; Restrepo et al., 2006; Meade et al., 1990; Govinda, 1993; YRCC, 2002; Sarma, 2005; Bassoulet et al., 1986)

of meters to kilometers long). The term 5-meander used herein, is based on splitting the total profile into reaches based on a river bending back and forth on itself five times (Figure 2.2). The longest reach, the profile, is the total gradient over the entire longitudinal profile. The sinuosity changes over the lengths of these river reaches, and is measured for each scale. Focusing on the river floodplain up to 100 meters is an arbitrary mark meant to analyze the lower most portion of a river system (Syvitski et al., 2009).

Varying the scale, or length of reach, to analyze the results of slope vs. sinuosity are performed to test the theory that the reach length examined “should be at least 20 times the average width of the channel” (Bell and Vorst, 1981). This scale dependency on slope and sinuosity are examined to determine which scale rivers can be compared to other systems, and at which scale, if any, inter-basin slope and sinuosity can be understood with relation to constraints on the system.

Creating plots of slope vs. sinuosity, following the flume experiment carried out by Schumm and Khan (1972), provides a much needed update of understanding rivers’ slope
and sinuosity in relation to other examined river systems (Figure 2.1). Gradient changes, appearing in the high resolution DEM as areas of steepening or flattening of the water surface, are overlaid on available geology and fault maps to assess controls and constraints applied to the river system.

Figure 2.1 After Schumm and Khan 1972, illustrating flume results of varying slope to address changes in channel patterns.

River longitudinal profiles are central to drainage basin geomorphology, and together with the channel system fix the boundary conditions in which hill slope processes occur (Knighton, 1998). The longitudinal profile of a river is defined as a function of its height ($H$) against the distance travelled downstream ($L$)

$$H = f(L)$$

Eq.1

The longitudinal profile of a river path is determined from a digital elevation model (DEM) where the flow path of a river is determined by water boundary data. Along with assistance from Germany and Italy, NASA and NGA funded the Shuttle Radar Topography Mission (SRTM) in 2001. The mission produced an almost worldwide DEM,
which provides relevant data for this study. The coverage includes the land area between 54 South to 60 degrees North with a resolution of 3 arc-seconds (~90 meter). The vertical resolution is 1 meter and referenced to WGS 84, with a vertical error of less than 10 meters for flat surfaces (Farr et al., 2007). The vertical precision of the SRTM (C-band) data depends on terrain characteristics, location on the globe and the feature of interest, and has a relative global accuracy of ≈ 3.7 m (Berry et al. 2007); the world’s flatter lying areas have a vertical RMSE (Root Mean Square Error) between 1.1 to 1.6 m (Schumann et al. 2008). Radar or more specifically Interferometric Synthetic Aperture Radar (InSAR) was used for the SRTM mission. Radar is not capable of penetrating the water surface, this allows for the analyses of water elevations and their resulting slopes downstream (Herrera-Cruz and Koudogbo, 2009). This incapability of penetrating the water surface is due to the river’s water surface being turbulent. Turbulence roughens the surface, creating a diffuse surface for the incoming radar signal, and allows for a returned signal to the satellite (Herrera Cruz and Koudogbo, 2009). Slater et al. (2006) determined the river elevations by sampling the water elevations perpendicularly from the centerline of the river network out to the neighboring riverbanks. Slater et al. (2006) applied an “iterative fitting mechanism”, using the average of these samples stored at each point to cancel erroneous elevation resulting from errors or correlated noise from the river. A graph of estimated elevation versus the centerline was generated and used to calculate an overall robust monotonic function (Slater et al., 2006). This process resulted in the delineation and creation of the elevations used in this study to extract the longitudinal profile.

The Surface Water Body Dataset (SWBD), a composition of Landsat images, was developed for use with the SRTM dataset for delineating large water bodies. The SWBD
outlines ocean bodies, major lake features (if > 600m length and >183m width) and major river features (if > 183m width for > 600m length, and ends after river width < 90m for < 1km). This dataset is in vector format, geographic coordinates WGS-84, and has a 90% confidence associated with its accuracy (Farr et al., 2007).

Figure 2.2 Identifying 1-meter sub-reach scale versus the 5-meander reach scale.

The SRTM and SWBD datasets are merged together, as they come in 1 by 1 degree tiles, with GIS tools, and projected into the corresponding Universal Transverse Mercator (UTM) zone for each river system. Utilizing the SWBD as a guide, a river centerline is created by hand digitizing (tracing) the center of the main river channel from the mouth up to the 100-meter WSE in the river. In locations where rivers split into two (or more)
channels, the widest, largest is assumed to be the main channel. Mid-channel islands or bars where ignored if their largest width is less than the combined width of the split channels. Segments of the centerline are split corresponding to the slope segments of the SRTM elevation data. Calculating the local elevation change over the longitudinal profile indicates channel gradient (slope) at the reach scale (Knighton, 1998). Slope \( S \) is given as the change in elevation \( \Delta H \) over the channel length \( C_L \).

\[
S = \frac{\Delta H}{C_L}
\]

Eq. 2

Concavity is calculated for all selected rivers. Concavity can be obtained via several methods, with the methods used for this study following those of Langbein (1964), Phillips and Lutz (2008), and Larue (2008). Termed concavity index, this method is computed based on deviations from a straight-line profile:

\[
CI = \sum \frac{(H_i^* - H_i)}{N}
\]

Eq. 3

Where \( H_i \) (meters) is the elevation at distance \( i \) (meters), and \( H_i^* \) (meters) is the elevation along a straight line from the uppermost to lowermost point along a river profile at horizontal distance \( i \), and \( N \) being the total number of measurement points (Langbein, 1964; Phillips and Lutz; 2008). \( CI \) values greater than 0 indicate concavity, and negative values indicate convexity (Langbein, 1964; Phillips and Lutz; 2008). Calculating the relative concavity for a profile with a given maximum elevation \( H_{max} \) is:
\[ CI_{relative} = \frac{CI}{(H_{max}/2)} \]  

Eq.4

\( CI_{relative} \) varies between -1 and 1, with 0 indicating a straight line (Langbein, 1964; Phillips and Lutz; 2008).

Utilizing the centerline, sinuosity is calculated for each channel profile segment. The sinuosity \( (Si) \) of a river segment is calculated by:

\[ Si = \frac{C_L}{V_L} \]  

Eq.5

Where \( C_L \) is the channel length and \( V_L \) is the straight-line valley length (Figure 2.3).

The channel length following the river from point A to point B, divided by the straight-shot distance from point A to point B. All values are \( \geq 1 \). Sinuosity provides a secondary basis for classification of the river (Knighton, 1998).

Figure 2.3 Sinuosity is calculated as the channel length divided by the straight-line valley length.
The longitudinal profile, and its derivative of slope and secondary classification of sinuosity, is examined by overlay analysis with auxiliary datasets, such as the SWBD vector files containing river confluences, and available geologic and tectonic maps. Steepening in the longitudinal profile is analyzed to determine if a nearby convergent channel is evident. This convergent channel could effectively increase the discharge or sediment load. The geologic and tectonic maps provide evidence of a change in underlying bedrock type or a point where the river contacts a fault.

For rivers located in the United States USGS Generalized Geological data is available and it is possible to overlay the floodplain maps with the underlying bedrock type, as well as fault lines for the entire country (Reed and Bush, 2004). Specifically for the Lower Mississippi River, a map locating regions of uplift is digitized and recorded to an accuracy of ~215 meter RMSE (Harmar and Clifford, 2007). Similarly, the Columbian Institute Ingeominas offers a digital Geologic Map of Colombia with bedrock type and fault lines (Gomez et al., 2007), which were overlaid with the Magdalena River floodplain map. The Colombian Geologic Map is digitized to high accuracy and ~130 meter RMSE with 16 tie points via geo-referencing.

Each longitudinal profile is marked accordingly with possible causes of convexities or concavities associated with changes in discharge, sediment load, lithology, or faults. This research examines real world rivers, therefore the final results are applicable to real systems, furthering knowledge of how rivers in the natural world shape and create the fluvial landscape. By including a range of fluvial environments, rivers can be compared,
highlighting their similarities and differences. Slope, climate zone, and lithology are analyzed to determine if systems can be grouped in these categories, allowing for an understanding of what controls place rivers into the three planform classifications set forth by Leopold and Wolman in 1957.

Common bedload equations include Meyer-Peter and Müller (1948), Einstein (1950), Bagnold (1980), and Ackers and White (1973). Bedload transport can be determined from three different classes of theoretical-empirical formulas, those that are based on 1) mean tractive forces 2) the statistical considerations of lift forces (a result of stream line convergence) 3) river slope and discharge relationships (Syvitski and Alcott, 1995). The study herein employs the third method; more specifically bedload $Q_b$ (kg/s) is calculated using a modified Bagnold (1966) equation

$$Q_b = \frac{\rho_s}{(\rho_s - \rho)} \frac{\rho g Q^\beta S e_b}{g \tan \lambda}$$

Eq.6

where $\rho_s$ and $\rho$ are the densities (kg/m$^3$) of sediment and water respectively, $g$ is the acceleration due to gravity (m/s$^2$), $S$ is the river’s gradient (m/m), measured from the SRTM DEM along the main channel, $Q$ is the local discharge, $e_b$ is the bedload efficiency (dimensionless), $\beta$ is a bedload rating term (dimensionless and set to 1 for this study), and $\lambda$ is the limiting angle of repose of sediment grains lying on the river bed (Syvitski and Alcott; 1995). Rivers in the presented floodplain database are relatively fine-grained, moving in sheets, and thus Bagnold's approach is sensible (Syvitski and Alcott; 1995).
The SRTM data is a snapshot in time, a temporal slice of 2001. Using the SRTM as well as other time stamps 1944 (Toposheets) and 2010 (MODIS) both pre and post-flood, an understanding of lateral migration for the Indus River can be calculated. The 1944 toposheets are at a scale of 1:63,000 to 1:250,000 and of comparable resolution to the 90 m SRTM maps. They are compiled mainly from Survey of India Maps, with aerial photo updates prepared by Army Map Service (GPDE), Corps of Engineers, U.S. Army Washington DC (See list of Maps used at the end of Chapter 7).

Lateral migration is calculated every two kilometers by using the centerline of the 1944, 2001, and 2010 both pre and post-flood. Perpendicular lines are generated from the early date, earlier date of the two centerlines being compared, calculating the distance between the two centerlines. By taking the average of these segment lengths, the average distance of lateral migration can be determined.

The eighth chapter discusses oxbow lakes, both the creation and importance of these architectural parameters in meandering river floodplains. The distance from the river centerline is used to create a cumulative oxbow lake area curve away from the river. Using this technique, an understanding of the population of oxbow lakes throughout the floodplain can begin to be unraveled. Applying this technique on both the Fly and Mississippi Rivers allows for scaling relationships to be examined. This analysis relies upon the SWBD, and the extraction of oxbow lake features from other water body features with the aid of Landsat.
Chapter 3 Quantifying Longitudinal Profiles of Floodplains

To regard the longitudinal profile as an ideal graded form, the initial problem is to determine the possible form that this function might take. Hack (1957) argued that slope and distance are related by a power function resulting in the understanding that a river’s profile could be explained through either a power or a logarithmic function. Snow and Slingerland (1987) showed that exponential, power, or logarithmic functions can all provide a reasonable fit to river profiles. They show the best fit function depends on which of the river’s controlling variables is exerting a dominant influence on the longitudinal profile. Power functions characterize downstream increase in sediment and water discharge; logarithmic functions characterize a rapid downstream decrease in bed material; and exponential functions characterize a decrease in transported material grain size under low sediment transport with all other major controls (sediment and water discharge) remaining constant (Knighton 1998). The three (power, log, and exponent) functional forms all provide smooth, concave-upward curvature.

The longitudinal profile of all 16 rivers are obtained by mapping the centerline of each river, followed by assigning each river segment the corresponding elevation from the SRTM DEM with discrete elevation steps of 1 meter, due to the vertical resolution of the SRTM data (Figure 1.1). All the rivers are mapped from the river mouth at sea level, 0 meters up to the 100-meter contour. The Eel, Kikori, Kapuas, Barito, Guaviare, and Kura rivers are unable to be mapped to their 100-meter contour or WSE due to their limited width in their upstream segments. For those systems the river channel width becomes less
than the resolution required (180 meters) to continue mapping the river’s change in
elevation upstream. Those six rivers are mapped to the highest elevation the SRTM DEM
resolution allowed. Three other rivers, the Niger, Columbia and Volga Rivers, are not
mapped up to the 100-meter contour of the WSE due to human interference in their basin.
Large dams are located in the main channel of the lower reaches of these rivers, and these
dams are so high that directly upstream of the dam elevations the water is higher than the
100-meter mark. Therefore these rivers are only mapped until the downstream terminal
end of their dam. All seven remaining rivers are mapped, overcoming resolution issues
for some (Indus and Kikori rivers). Again, these resolution issues are due to a region of
the river where the river’s width is too small to be fully resolved by the SRTM data. This
is overcome by interpolating the WSE that is captured above and below the river sections
with a small width. In order to represent all 16 of the longitudinal profiles on the same
chart the channel segment lengths are normalized (Figure 3.1). Each river shares a
similarly unsmooth and deviant profile, some with large steps and others with smaller
more gradual slope changes.

A trend fitting comparison is performed to compare the rivers quantitatively against one
another (Snow and Slingerland, 1987; Knighton, 1998; Harmar and Clifford, 2007).
Regression techniques are applied to each longitudinal profile to find the best
mathematical fit. Following Snow and Slingerland (1987), a power function is the best fit
for the majority of the rivers. A power function characterizes downstream increase in
sediment and water discharge (Snow and Slingerland, 1987; Knighton, 1998). The
Magdalena River’s longitudinal profile showed a best fit with an exponential function.
Figure 3.1 Normalized longitudinal profiles of the 16 selected rivers.

Figure 3.2 Normalized, to total river length, longitudinal profiles of the 16 selected rivers.
This is also the case for the Guaviare River. Exponential functions characterize decrease in grain size under low sediment transport with all other major controls kept constant (Snow and Slingerland, 1987; Knighton, 1998). Only one river, the Kapuas River, is best represented by a log function, which characterize a rapid downstream decrease in bed material (Snow and Slingerland, 1987; Knighton, 1998). These results indicate that the selected rivers represent a variety of fluvial systems with differing longitudinal profiles. However, this analysis only examines the lower portion of the respective longitudinal profile, i.e. the lowermost floodplains and deltaic floodplain.

To determine the fraction of a river below the 100-meter contour in respect to the total length of the main channel, the total length of the main channel of each river is determined from literature. Figure 3.2 presents the normalized length of each river examined to the total river length and reveals that mapping to the WSE 100 meter contour provides a different percentage of each river mapped. Examining the Figure 3.2 it is decided to compare rivers in which 30 percent or more of the river’s WSE is mapped. The Mississippi, Magdalena, Mahanadi, Indus, Brahmaputra, Fly, Kikori, Barito, Kapuas, and Guaviare longitudinal river profiles are again regressed to determine the best fit. This time all rivers showed best fit with a power function, again characterizing downstream increase in sediment and water discharge (Snow and Slingerland, 1987; Knighton, 1998). The Mississippi, Mahanadi, Indus, Brahmaputra, Fly, and Kikori experienced a ≤0.02 change in the root mean square from the previous regression fit. The Barito experienced a slightly larger change of 0.05 in root mean square. The Magdalena, Guaviare, and Kapuas rivers (previously fitted with exponential and log trends) experienced the largest changes
in best fit (0.15, 0.13, 0.07 respectively). This illustrates that for these ten rivers with their longitudinal profiles mapped at 30% or more, is governed by a downstream increase in sediment and water discharge, analogous to the theory behind that of a power trend (Snow and Slingerland, 1987; Knighton, 1998).

Secondly, the lower most portion of a river system, the deltaic region, is in part a bypass system where only a portion of the sediment that travels down the river is retained (Overeem et al., 2005). Following this, the evolutionary process of a river needs to be considered (Ohmori, 1991). Although river studies are accustomed to trend fitting with exponential, log, and power functions; rivers can also be matched by linear functions, indicating that their longitudinal profiles represent almost straight lines (Ohmori, 1991). A linear fit would theoretically indicate that the river is in mass equilibrium: the sediment load is balanced between inflow and outflow (Ohmori, 1991). Through aggradational processes, the shape of a river’s longitudinal profile changes with the change in mathematical function type, from exponential, to power, and finally to linear functions (Ohmori, 1991). It is necessary to fit the simplest mathematical model to a series of points. This allows for the easiest and most straightforward understanding of the phenomena being observed. A linear fit is also applied in this study to the 16 rivers depicted as the focus of this study is on the lowest most portion of the fluvial channel. A linear trend is appropriate for the lowermost part of a river system (i.e. delta) as this is the depositional part of the longitudinal profile (Ohmori, 1991).

In general the systems in this study are well represented by linear functions, with an
average $R^2$ of 0.95 ±0.05 (Table 3.1). Systems with a slightly smaller $R^2$ are the Columbia, Magdalena, Kapuas, and Kikori (0.85, 0.88, 0.89, and 0.89 respectively) and are associated with large-scale knickpoints. These large-scale knickpoints deviate significantly from the linear fit function, causing a slightly lower $R^2$ than the other twelve systems. Following previous rationale of comparing the river’s with an equal percentage (30) of river’s WSE mapped, the Mississippi, Magdalena, Mahanadi, Indus, Brahmaputra, Fly, Kikori, Barito, Kapuas, and Guaviare longitudinal river profiles are also regressed to linear functions. The Brahmaputra, Indus, Mississippi, Fly, Kapuas, and Kikori Rivers all reflected a higher $R^2$ for the lower 30% of river, compared to the total river examined. The Magdalena, Mahanadi, Barito, and Guaviare Rivers experienced a lower $R^2$ for the lower 30% of the river, compared to the total river examined. Comparing these 10 fluvial system’s lower 30% indicates that there is not a tendency for a river to become better represented with a linear function in its lower reaches. Following theory, the Brahmaputra, Indus, Mississippi, Fly, Kapuas, and Kikori Rivers indicate a greater equilibrium of mass (the sediment load is balanced between inflow and outflow) for their lowest 30% (Ohmori, 1991). The Magdalena, Mahanadi, Barito, and Guaviare indicate a greater equilibrium of mass at their full examined length, lesser so at the lowest 30%; indicating these four systems are more erosional or depositional in their lower reaches than the total river longitudinal profile examined for this study.

The fact that the systems in this study are well represented by linear functions, with an average $R^2$ of 0.95 ±0.05 could be seen as problematic. The results from Syvitski et al. (2009) found that many deltas are sinking partly due to the loss of sediment to the delta
along with other factors. If these lowermost portions of the river systems are linear in their lower reaches, instead of concave up, this deltaic region is indeed a bypass system where only a portion of the sediment that travels down the river is retained (Ohmori, 1991; Overeem et al., 2005). This metric can be further used to determine which deltas, already determined at great risk or great peril by Syvitski et al. (2009), are heading to collapse in their near geologic future.

<table>
<thead>
<tr>
<th>Table 3.1</th>
<th>Total River Length</th>
<th>Length Examined</th>
<th>Fraction Power R²</th>
<th>Linear R²</th>
<th>CI-relative</th>
<th>Linear (30%) Power</th>
<th>CI-relative (30%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Brahmaputra</td>
<td>2900</td>
<td>1085</td>
<td>0.37</td>
<td>0.97</td>
<td>0.98</td>
<td>0.37</td>
<td>0.99</td>
</tr>
<tr>
<td>Indus</td>
<td>3180</td>
<td>1139</td>
<td>0.36</td>
<td>0.96</td>
<td>0.99</td>
<td>0.24</td>
<td>0.99</td>
</tr>
<tr>
<td>Magdalena</td>
<td>1540</td>
<td>755</td>
<td>0.49</td>
<td>0.96*</td>
<td>0.88</td>
<td>0.38</td>
<td>0.82</td>
</tr>
<tr>
<td>Mahanadi</td>
<td>900</td>
<td>357</td>
<td>0.40</td>
<td>0.98</td>
<td>0.99</td>
<td>0.20</td>
<td>0.97</td>
</tr>
<tr>
<td>Mississippi</td>
<td>3730</td>
<td>1697</td>
<td>0.46</td>
<td>0.98</td>
<td>0.96</td>
<td>0.20</td>
<td>0.98</td>
</tr>
<tr>
<td>Niger</td>
<td>4180</td>
<td>824</td>
<td>0.20</td>
<td>0.99</td>
<td>0.98</td>
<td>0.23</td>
<td>-</td>
</tr>
<tr>
<td>Yellow</td>
<td>5464</td>
<td>825</td>
<td>0.15</td>
<td>0.99</td>
<td>0.98</td>
<td>0.16</td>
<td>-</td>
</tr>
<tr>
<td>Fly</td>
<td>1050</td>
<td>860</td>
<td>0.82</td>
<td>0.99</td>
<td>0.92</td>
<td>0.50</td>
<td>0.95</td>
</tr>
<tr>
<td>Kura</td>
<td>1515</td>
<td>793</td>
<td>0.52</td>
<td>0.92</td>
<td>0.97</td>
<td>0.15</td>
<td>-</td>
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<tr>
<td>Volga</td>
<td>3692</td>
<td>637</td>
<td>0.17</td>
<td>0.98</td>
<td>0.94</td>
<td>0.44</td>
<td>-</td>
</tr>
<tr>
<td>Eel</td>
<td>320</td>
<td>71</td>
<td>0.22</td>
<td>0.92</td>
<td>0.97</td>
<td>0.06</td>
<td>-</td>
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<tr>
<td>Columbia</td>
<td>1953</td>
<td>451</td>
<td>0.23</td>
<td>0.90</td>
<td>0.85</td>
<td>0.46</td>
<td>-</td>
</tr>
<tr>
<td>Barito</td>
<td>890</td>
<td>635</td>
<td>0.71</td>
<td>0.91</td>
<td>0.98</td>
<td>0.23</td>
<td>0.94</td>
</tr>
<tr>
<td>Kapuas</td>
<td>1143</td>
<td>879</td>
<td>0.77</td>
<td>0.96*</td>
<td>0.89</td>
<td>-0.09</td>
<td>0.91</td>
</tr>
<tr>
<td>Kikori</td>
<td>320</td>
<td>184</td>
<td>0.57</td>
<td>0.96</td>
<td>0.89</td>
<td>0.27</td>
<td>0.93</td>
</tr>
<tr>
<td>Guaviare</td>
<td>1497</td>
<td>1170</td>
<td>0.78</td>
<td>0.99*</td>
<td>0.98</td>
<td>-0.51</td>
<td>0.89</td>
</tr>
</tbody>
</table>

*Magdalena and Guaviare – Exponential, Kapuas -Logarithmic

Thirdly, concavity of the longitudinal profiles is analyzed. Following equilibrium theory, profiles should be concave up (Hack, 1957; Snow and Slingerland, 1987; Knighton, 1998; Harmar and Clifford, 2007). Comparing concavity of longitudinal profiles indicates the river’s steady state equilibrium, assuming that smooth concave longitudinal profile is an expression of steady state equilibrium. Typically, older rivers are more concave as they
have had significantly more time to transition to graded equilibrium (Knighton, 1998).

Applying the relative Concavity Index (Eq.4) on the sixteen rivers shows that fourteen of the rivers are slightly-concave to concave, while the Kapuas and Guaviare rivers are slightly-convex and convex respectively. The convexities could be due to only examining the lower portion of the rivers, but each of these rivers has over 70% of the river WSE mapped. The anomalously strong convexity, -0.51, of the Guaviare can be explained due to the large increase in slope as the river merges with the Orinoco. The Guaviare River is a tributary to the much larger Orinoco River. The large erosional power and sediment yield of the Orinoco, with an average discharge of 34,500 m$^3$/s and sediment yield of 180 T/km$^2$/yr, has resulted in lowering of the base level of the Guaviare outflow (Syvitski and Saito, 2007). This base level lowering due to the influence of the Orinoco River is the most apparent cause in the large slope in the lowermost reach of the Guaviare River. The Guaviare River cannot keep pace with the rapid incision of the Orinoco River. The slight convexity of the Kapuas River is due to a bedrock outcrop of Borneo in which this system has not had significant time to incise (Bruggen, 1955). On the paleo-course of the Kapuas River, Bruggen (1955) suggested that in the Quaternary the present Kapuas broke through the Semitau uplands to converge with the Melawi River near present day Sintang. The older course of the Kapuas River used to flow northward to the Kapuas Lakes area, following the course of the present Batang Lupar River (Bruggen, 1955).

Following previous rationale of comparing just the rivers with an equal percentage (30%) of their WSE mapped, the Mississippi, Magdalena, Mahanadi, Indus, Brahmaputra, Fly,
Kikori, Barito, Kapuas, and Guaviare longitudinal river profiles are calculated using the relative Concavity Index (Eq.4). The Brahmaputra, Indus, Magdalena, Fly, and Kikori Rivers experienced a decrease in the relative Concavity Index, or a transition towards linearity (Table 3.1). The Mahanadi, Mississippi, and Barito Rivers experienced an increase in the relative Concavity Index, or more concave for the lower reaches (Table 3.1). The Kapuas transitioned from slightly convex to slightly concave, whereas the Guaviare transitioned from convex to slightly convex.

The longitudinal profile of the Fly River is strongly concave (Relative Concavity Index: 0.50). This is attributed to the fact that the Fly River is the only river where 80% of its total longitudinal profile is below the 100-meter contour. Typically, the concavity index is higher as a larger portion of a river’s longitudinal profile is examined (Hack, 1957; Snow and Slingerland, 1987; Knighton, 1998; Harmar and Clifford, 2007). The study generally corroborates that concavity increase with increasing inclusion of the drainage basin uplands, following both theory and empirical data (Table 3.1). However, there are a few exceptions, the Guaviare and Kapuas Rivers are mapped to 78 and 77 percent of their total length respectively, and both of these systems had a negative relative Concavity Index. Concavity appears to not hold at the lower reach scale examined in this study. In fact, previous studies have expressed that while theoretical possibility of obtaining a graded, concave longitudinal profile exists, it is not reasonable to expect (Phillips and Lutz, 2008) as some rivers, including some alluvial rivers, are not capable of adjusting gradients to achieve concavity (due to underlying geology or anthropogenic influences; Xu, 1991). Another issue is termed equifinality, since different constraints and processes
can produce the same effect of a smoothly concave longitudinal profile (Snow and Slingerland, 1987; Ohmori, 1991; Sinha and Parker, 1996: Whipple, 2004; Harmar and Clifford, 2007), including non-steady state conditions. Since non-steady state processes can produce a smooth profile, the presence of a smooth concave profile is not necessarily indication of grade or equilibrium state (Phillips and Lutz, 2008).

A study of the Mississippi River’s lower longitudinal profile by Harmar and Clifford (2007) illustrates the problematic nature of attempting to apply concave profiles as indication to determine if a river is in equilibrium (Phillips and Lutz, 2008). Harmar and Clifford (2007) found that although the Mississippi River longitudinal profile is concave at the largest scale, it is characterized by shorter trends, zonal variations, and discontinuities. These local deviations from the overall concave profile are a response to bed material changes and morphology related to physical and engineering controls (Harmar and Clifford, 2007). The second most concave river with a Relative Concavity Index of 0.46 is the Columbia River. Sherwood and Creager (1990) found that the Columbia River’s longitudinal profile has probably not, attained a final equilibrium with modern conditions because tide range, river flow, sediment supply, and human factors are constantly changing.

The longitudinal profiles of rivers found in nature are almost never smooth but tend to be concave-up, often containing local convexities (Knighton, 1998). The variations of steepening/flattening seen in a longitudinal profile imply that convexities are abnormal, where a smooth concave-up profile is diagnostic of the graded or equilibrium state river
Local steepening of a longitudinal profile can result from one of several causes: “a) more resistant bedrock strata; b) the introduction of a coarser or larger load; c) tectonic activity or d) the effect of past events, notably fall in base level” (Knighton, 1998). Schumm, Leopold and Bull (1993; 1979) suggest that the effects of base-level (sea level) change are not transmitted for long distances upstream, especially in large alluvial systems. Large alluvial systems have the ability to absorb change by adjusting their sinuosity, channel dimensions and roughness characteristics (Knighton, 1998). Slope change can be initiated with a fall in base level, while bedrock resistance is the controlling factor that determines the rate at which steepening occurs (Brush, 1961). Bedrock in the fluvial systems can even constrict lateral movement of a river leaving vertical adjustment the sole option (Knighton, 1998).

For this study concavity is analyzed based on regression analysis to evaluate the relationship between concavity and slope or sinuosity. Following previous studies concavity and slope have no relationship ($R^2 < 0.05$) (Phillips and Lutz, 2008). Concavity and sinuosity also have no significant relationship ($R^2 \sim 0.09$).
Chapter 4 Floodplain Channel Slope and Sinuosity

Channel patterns are representations of channel form adjustment in the plan-view, and linked to lengthwise modes (Knighton, 1998). Channel pattern influences the resistance to flow and can be regarded as a method of slope adjustment with valley slope held constant (Knighton, 1998). The effect of a meander is to increase a river’s resistance to flow, effectively reducing the channel gradient (Knighton, 1998). This is intuitive since a straight line is the most ordered state a river can take (perfect axial symmetry and zero entropy) (Stolum, 1996).

4.1 Slope and Sinuosity over the Entire Profile

Comparing the entire longitudinal profile of each river through the variables of slope and sinuosity, significant relationships are established (Figure 4.1.1). The results indicate a high correlation between slope and sinuosity with 56% of the variance explained. Eliminating the furthest outlier, the anastomosing Volga River (major human influence along its course) increases the correlation between slope and sinuosity, explaining 75% of the variance. This strongly indicates that slope has a moderate to high level of control on river channel sinuosity.

An important early planform classification by Leopold and Wolman (1957) divided rivers into three categories: braided, meandering and straight. The absence of long straight reaches and the presence of sinuous flow in straight reaches are regarded as evidence of an
Fig 4.1.1 The results of the Entire Longitudinal Profile indicate a high correlation between slope and sinuosity with 56% of the variance explained.

inherent tendency of natural streams to meander, respective of scale or boundary material. Any river stretch with a sinuosity $>1$ is seen as a sinuous river (Knighton, 1998). For this study the sinuous rivers are examined and are evaluated at values less than a sinuosity value of 1.5, as sinuosity of 1.5 is often used as the threshold to obtain the classification of meandering (Knighton, 1998). Analyses indicate that the straightest rivers (sinuosity values $<1.4$) have the most variable slope values. Rivers with the highest slope are draining areas of active orogenic margins (Cascade Mountains, Columbia and Eel), recent volcanic activity (Central Range, Kikori), and relic folded margins (Mahanadi Satpura Brahmani Range) (Coleman et al., 2008). The sinuosity of the Yellow River is surprisingly low for the relatively low slope but is artificially constrained by hardened levees (Figure 1.4). Combining this with a steepened riverbed gradient (2.9 times) due to
human intervention via reduced river length illustrates that the Yellow River is experiencing disequilibrium and scouring (Hui and Huang, 2005). The Volga River is the largest outlier from the regression as expressed earlier. This is believed to be due to the low slope of the system’s subaerial and subaqueous delta combined with the effects of the highly variable Caspian sea-level (Kroonenberg et al., 1997).

As previously indicated, sinuosity values of 1.5 is seen as the threshold in which a sinuous river becomes meandering. Based on the values of sinuosity calculated for the entire river’s longitudinal profile for all sixteen fluvial systems, the threshold is raised to 1.6 as discussed in the next paragraph. Three river systems (the Brahmaputra, Magdalena, and Niger Rivers) are found to have relatively high sinuosities with values ranging between 1.4 and 1.55. The Brahmaputra and Magdalena are draining active orogenic regions, the Himalaya and Ande Mountain Ranges respectively. The Brahmaputra has a classic braided form, whereas the Magdalena is more of an anastomosing river. These river channels carry large influxes of sediment due to high rates of erosion, with sediment yields of 701 and 868 T/km²/yr respectively (Syvitski and Saito, 2007). The Niger River has a relatively high sinuosity, which is due in most part to its highly sinuous stretch on the delta where the river branches into >25 mouth outlets, as counted from SWBD. The two largest distributary channels of the Niger River have high sinuosity.

The most intriguing threshold found in this study is established using the sinuosity classification. A sinuosity value of 1.6 illustrates the lower limit in which a meandering system is found, using the entire river longitudinal profile. This sinuosity value, at the
scale of the entire river’s longitudinal profile, is the threshold at which a meandering system is attained. The meandering channel remains an important end-member in all modern classifications of river planform and one of the most common patterns found in nature (Hickin, 2003). The Indus and Mississippi Rivers, both large-scale fluvial systems, barely exceed this threshold and contain slopes just below 10 cm/km. The remaining five Rivers: Kapuas, Kura, Barito, Guaviare, and Fly, all have a sinuosity value greater than 1.9 (Figure 4.1.1). The meandering fluvial threshold of 1.6 is not only evident in high values of sinuosity, but is also apparent in the floodplain architecture.

Growth and cutoff of meanders are important in the construction of the meandering river’s floodplain (Wolman and Leopold, 1957), and its alluvial architecture (Allen, 1965; Bridge et al., 1986; Constantine and Dunne, 2008). All seven of the meandering rivers are actively producing oxbow lakes, evident in the oxbow lakes along the longitudinal stretches examined and found throughout the floodplain. Oxbow lakes are the products of meander cutoff, and their occurrence and sedimentary deposits influence rates of meander migration (Hudson and Kesel, 2000; Constantine and Dunne, 2008). The process of lateral migration and the production of oxbow lakes are the two main processes responsible for development of sand bodies (Richardson et al., 1987). Over geologic time rivers shift and these sand bodies slowly fill in with deposit sheets of silt and clay, can resulting in a reservoir capable of sequestering hydrocarbons (Richardson et al., 1987). Predictive relationships of oxbow lake production and lateral migration rates are of great interest to the field of Sedimentology (Richardson et al., 1987). To assess the ability to use data sets and methods herein described on the determination of lateral migration rates
(for the Indus) and the power of predictive relationships for oxbow lake production (for the Fly and Mississippi), see Chapters 7 and 8 respectively.

Previous studies indicate that examining the slope and sinuosity of a river over an extended length is a common method of comparing Rivers to one another, it is also crucial to understand a River’s internal relationships (Harmar and Clifford, 2007). To downscale the scope of a river examined from the entire longitudinal profile to a reach scale, a 5-meander “scale” is selected. This allows for segmenting the entire longitudinal profile into reaches in which to analyze the river’s change in slope and sinuosity at different locations. Tidal dominated deltas are examined, as they are one end-member of Galloway’s (1975) classification.

4.2 5-meander Reach Scale

The 5-meander reach scale slope vs. sinuosity is mapped with the entire longitudinal profile slope vs. sinuosity in Figure 4.2.1. This allows the viewer to examine the variation of slope and sinuosity, at the 5-meander reach scale, against a larger averaged gradient and channel sinuosity of a longer stretch of river. The anomalously high slope or sinuosity at the 5-meander reach scale is seen as large deviations from the mean of the entire longitudinal profile. This allows for pulling apart of local river reaches to understand outside influences on the fluvial system (i.e. tides).
Figure 4.2.1 The Entire Longitudinal Profile’s slope and sinuosity shown with the 5-meander reaches.

### 4.2.1 Tidal

Most rivers lack gauging stations disallowing tidal limits to be mapped, therefore the tidal limits are determined from literature. The Fly River’s tidal effects are felt 100 km upstream of the delta to the end of the fluvial reach (EFR), and during low-flow periods tidal effects can be felt up to Manda (which is above midway of the middle Fly) (Lauer et al., 2008). Local tides range from 3.5 m at the mouth to 5 m at the apex (Wolanski and Eagle, 1991). Using the EFR as the boundary, sinuosity of the main channel is 1.1 in the tidal domain and 2.6 upstream of the tidal limit. Using Manda as the boundary gives a higher sinuosity of 1.7 below the tidal limit, and no change for the sinuosity upstream of the tidal limit. The Kikori River’s tidal range is about the same as the Fly’s, as they both...
drain into the Gulf of Papua ~100 km apart, and shows a sinuosity of 1.2 in the tidal
domain and 1.4 upstream of the EFR.

The Indus river feels tides reaching inward as far as Tott ~160 km upstream (Modern day
Thatta; Eisma 1998). The average tidal range is between 2 and 3 meters (Zaigham, 2005). The
sinuosity of the tidal reach up to Thatta, as calculated from the Indus’ active channel
using the SRTM is 1.65 while the fluvial reach above is 1.81. In fact, after a
comprehensive review of sinuosities over the period between 1944 and 2010 (see Chapter
7), the Indus River’s main channel indicates that the sinuosity of the tidal reach is always
less than the fluvial reach upstream of Tott.

In this study the Columbia River appears to be the only exception to the results thus far. The
tides entering the estuary are mixed diurnal and semidiurnal with a mean range of 2
meters (Sherwood and Creager, 1990). Tidally induced water fluctuations are felt as far
upstream as the Bonneville Dam ~220 km upstream (Clark and Snyder, 1969). The
calculated sinuosity for the tidal reach, below the dam, is 1.25 and 1.12 upstream of the
dam.

Tidal effects on the fluvial system appear to vary case by case. The Columbia River’s
sinuosity found to be greater in the tidal reach is the only fluvial system, out of the four
tidally dominated systems, to conclude with these results. The other three rivers: Kikori,
Fly, and Indus experienced sinuosities much greater in the fluvial dominated sections,
upstream of the tidal region. This could be explained by the fact that the Fly and Indus are
shown to be meandering systems, but the Kikori River is not. The one trend that holds true to all four systems is that the slope calculated in these tidally dominated reaches are considerably low compared to the fluvial reach.

4.2.2 Anthropogenic effects on Slope and Sinuosity of the Mississippi

Anthropogenic impacts on a fluvial system can be examined by utilizing 5-meander reaches of the longitudinal profiles. In the most widely documented engineering intervention performed on the Lower Mississippi River’s profile is the artificial shortening thereby removing 230 km of the planform by straightening 14 of the most sinuous bends between Old River and Memphis between 1930-43 (Harmar and Clifford, 2007; Moore, 1972; Winkley, 1977). The longitudinal profile of the Lower Mississippi has been artificially steepened, and bankline stabilizations since have removed the ability to lengthen the channel (Harmar and Clifford, 2007).

Adding the additional length of the artificially removed 230 km between Old River and Memphis causes the local sinuosity to jump to 2.04 from the current 1.36. This reveals that this stretch would have been ~1/2 as steep, and that the furthest most upstream reaches would have been steeper than the mid straightened reaches (Harmar and Clifford, 2007). The longitudinal profiles therefore show a more concave profile, which is consistent with a downstream increase in discharge. Adding this additional length back to the main channel, that the large engineering achievement removed, causes the natural river slope, of the entire ~1700 km examined, to decrease by ~1cm/km increasing the sinuosity to 1.95 from the current 1.69 (effectively shifting the system into the group with
the smaller meandering rivers (Figure 4.1.1)).

4.2.3 5-meander Slope and Sinuosity

The most intriguing results of the 5-meander reach scale study, is determined by visually comparing the spread of slope and sinuosity to a river’s entire longitudinal profile slope and sinuosity values. The spread of the 5-meander reaches generally encircle the entire river’s average slope and sinuosity value. This allows for the identification of values, beit slope or sinuosity, which are far removed from the entire river’s mean. For example, the one system that had no correlation between the 5-meander reach slope values and the entire longitudinal profile is the Columbia River. This is due to the numerous dams in the system, causing extremely low calculated slopes at the reach scale.

The result of comparing slope and sinuosity at the 5-meander reach scale indicate a low correlation with only 18% of the variance explained. This regression analysis shows that systems, as a whole, can be compared over a large-stretch averaged slope and sinuosity for each river. Analyzing the relationship of slope and sinuosity of a river by studying 5-meander reaches, however, weakens this relationship for each separate reach compared to the entire river slope-sinuosity relationship, but provides the possibility to study controlling processes in more detail. The 5-meander reach scale is a crucial measure in determining controlling factor as well calculating bedload (See Chapters 5 and 6). The SRTM vertical resolution is 1 meter, by using this as the smallest resolution scale (termed sub-reach herein), the river can be divided even further to analyze in more detail.
4.3 1-meter Sub-Reach Scale

Examining slope and sinuosity based on a 1 meter vertical measure pushes the envelop of the vertical error limits of the SRTM DEM. The high resolution SRTM DEM allows for determining local gradient change, local anomalies, and where the river steepened its descent on its course to base level. Each local sub-reach is established using the 1-m vertical break in the river’s water surface elevation (WSE) where the length of the sub-reach determines steepness; a very steep sub-reach would be comprised of a short segment. The sub-reach scale is important to analyze river systems with extremely localized events.

Discussed more thoroughly in the Controlling Factors Chapter, the 1-meter sub-reach scale is useful to determine locations with anomalously low or high slope and sinuosity values. The 1-meter sub-reach scale is specifically useful in determining the effects different fault types have when intersecting, at varying degrees, the main river channel. This technique also proves useful in analyzing the slope changes associated with transitions to and from bedrock and regions experiencing significant uplift or subsidence. To illustrate the slope and sinuosity differences, Figures 4.3.1 and 4.3.2 are the slope and sinuosity calculated at varying scales (5-meander reach and 1-m sub-reach) for the Mississippi and Niger Rivers (more on how this is used in Chapter 5).
Figure 4.3.1 Mississippi River slope vs. sinuosity at both 1-m sub-reach and 5-meander reach scales. The highest sinuosity measures are at the sub-reach scale and coincide with the rivers crossing growth fault scarps perpendicularly.

Figure 4.3.2 Niger River slope vs. sinuosity at both 1-m sub-reach and 5-meander reach scales. The highest sinuosity measures are at the sub-reach scale and coincide with the rivers crossing growth fault scarps perpendicularly.
Chapter 5 Neotectonics, Lithology and the Fluvial Response

The location and height of the start and end points of a longitudinal profile, and therefore its average gradient, are initially determined by tectonic and other geo-historical events (Knighton, 1998). Tectonics can occur over a range of scales affecting slope and supply of water and sediment to the basin (Knighton, 1998; Bridge, 2003). Periodic activity of a fault intersecting a river valley can result in local change in valley slope, river diversion, and change in channel pattern over a period of several hundred or thousand years (Schumm et al., 2000; Bridge, 2003). Tectonic activity can also result in a large topography change of the floodplain relative to the input of flow of water and sediment ultimately influencing sedimentation rates (Schumm et al., 2000; Burbank and Anderson, 2001; Bridge, 2003). Changes in both sinuosity and slope have been observed as a river crosses a fault (Lindanger et al., 2004; Lindanger et al., 2005). This chapter examines the 1-m sub-reach and 5-meander reach scales to determine the effect lithology and faults can have on a river channel.

Slope changes show a short-term (year to decade) response with a decrease in river sinuosity and braiding if slope is decreased, and an increase in river sinuosity and braiding where slope is increased (Schumm 1986; Holbrook and Schumm 1999; Schumm et al. 2000; Burbank and Anderson, 2001; Bridge, 2003). An uplifting ridge results in a decrease in slope upstream of the ridge, and an increase in the downstream slope assuming the axis is running along-valley; opposite fluvial responses occur in a subsiding basin (Bridge, 2003; Lindanger et al., 2005; Figure 5.1). Furthermore container valley widths are seen to have a huge impact on the sinuosity of a river’s planform sinuosity. Where a
container valley is narrow, or narrower than other parts of the river system, the river’s planform sinuosity is less than regions where a river is allowed to freely move about the alluvial terrace.

Figure 5.1 Illustrates the process locations associated under varying neotectonic processes.

5.1 Mississippi

There are two types of neotectonic processes affecting the Lower Mississippi. There are two regions of Uplift, the Lake County and Monroe Uplifts, and a seismic zone known as the New Madrid zone (Harmar and Clifford, 2007). These neotectonic processes can be seen in the Mississippi’s longitudinal profile, Figure 5.1.1, and the Map, Figure 5.1.2. Leveling surveys of both Uplifts conclude movements of ~4mm/yr (Burnett and Schumm, 1983; Gregory and Schumm, 1987). The Seismic Zone is reported to have produced the four largest earthquakes in eastern North America (1811-1812) in which large amounts of sediment entered the Mississippi due to large spread landslides and caving banks (Jibson et al., 1988; Harmar and Clifford, 2007).

Neotectonic uplift provides a physical constraint on a river system; it reduces the slope and stream power upstream of the uplift, simultaneously increasing the downstream slope
(Richards, 1987). The upstream end of the Monroe Uplift immediately follows the Arkansas confluence, and shows an increase in the elevation according to Hamar and Clifford (2007). This reach, due to the Monroe Uplift now consists of uplifted Tertiary bedrock, a non-alluvial hard stretch in the bed (Gregory and Schumm, 1987). Harmar and Clifford (2007) found no steepening at the downstream end of the Monroe Uplift, attributing this to the artificial cutoff program. The locations of the Lake County Uplift, and the New Madrid seismic zone also exhibit no formative geomorphological evidence according to Harmar and Clifford’s (2007) review of the Lower Mississippi’s bed profile. The SRTM data shows something different, as the data provides WSE, not the bed profile.

The Monroe Uplift shows a decrease in slope of the Mississippi at the upstream end of the resistant strata, and is followed by a slight increase in slope coinciding precisely with the downstream end of the uplift (according to the 1 m sub-reach and 5-meander reach measures). The slope of the 5-meander stretch at the downstream end of the Ohio River confluence is the highest of all stretches and is overlaying the Lake County Uplift (refer to Mississippi profile and Tectonics Map in Figures 5.1.1 and 5.1.2). The increased slope could be attributed to the increase in water and sediment discharge of the confluence, or could be a result of the increased gradient shedding the increase in sediment flux due to seismic events. It could also be from the uplift, an up-valley increase in supply rate and grain size of sediment (due to tectonics) with channel-forming discharge remaining constant necessitating an increase in slope (Bridge, 2003). The more interesting measure is the anomalously high sinuosity, because the increased slope can be accomplished short-term with a decrease in channel sinuosity and increase in degree of braiding, which is not observed at this zone (Bridge, 2003). This high slope and sinuosity could be due to the
uplift rates causing a localized geomorphological relic. The uplifts influence the longitudinal profiles at both the lowest 1 m sub-reach measures, and at the 5-meander reaches.

Figure 5.1.1 Mississippi River’s Longitudinal Profile.

A meandering river responds to a change in the gradient caused by a fault scarp, either straightening the river (lowering its sinuosity) with a gradient drop or increasing the sinuosity with an increase in gradient (Lindanger et al., 2004; Lindanger et al., 2005). Rivers are theorized and observed to respond to the subtlest change in underlying tectonics (Lindanger et al., 2004; Lindanger et al., 2005). This small change seen in the river system can be used to map deformational changes in the underlying strata (i.e. fault lines) (Bridge and Demicco, 2008). Growth faults are normal faults with the footwall on
the upstream end, and a hanging-wall experiencing the downward movement on the downstream end. Lindanger et al. (2004; 2005) examined growth faults of rivers, including the Mississippi, near Baton Rouge observing that rivers crossing scarps at a nearly perpendicular angle show a change in sinuosity, with a peak in sinuosity immediately following the intersection of the river and fault. Following that, a meandering river should increase its sinuosity with an increase in gradient.

Figure 5.1.2 Neotectonic Map of the Mississippi’s alluvial basin.
The Mississippi and Niger profiles both display an increased sinuosity on their deltas, as the rivers cross the scarps at a perpendicular angle, where they are experiencing growth faults (See faults in Figure 5.1.3 for the Mississippi and in Figure 5.1.4 for the Niger). The results of the increased sinuosity just downstream of the growth fault match the results of Lindanger et al. (2004; 2005). Figure 5.1.5 illustrates how a growth fault slightly increases the gradient over time causing the river’s energy to increase slightly, thereby increasing the sinuosity. The sinuosity immediately increases as its flows over the Niger delta, where there is experiencing growth faults throughout the delta. In examining another channel on the Niger delta, the sinuosity also is very high following the trend of the other main channel. The Mississippi delta has a much more variable location of growth faults. Located upstream of Baton Rouge, a series of growth faults are located perpendicular to the flow of the Mississippi. The river’s sinuosity increases and then straightens as if orients the flow parallel with a fault, visible in Figure 5.1.3. Although there is not an increase in gradient associated with this, the high sinuosity values are not associated with the most minimum gradients for either river.

The scale at which the geomorphological expression of the rivers crossing the fault scarps appears at the highest resolution of the SRTM DEM (being 1-m vertical measurement of sub-reach scale). Refer to the slope vs. sinuosity plots in Figures 4.3.1 and 4.3.2, where it is apparent in both systems that sinuosity is highest in longitudinal profiles at regions where the rivers’ flow is perpendicular to growth fault scarps. While the Mississippi sinuosity at 5-meander reach scale appears to muffle the fault’s existence, this could be due to the fact that the Mississippi is a highly sinuous system (Figure 4.3.1). The Niger’s sinuosity is still extremely high at this 5-meander reach scale (Figure 4.3.2).
Figure 5.1.3 Growth Faults seen on Mississippi Delta.

Figure 5.1.4 Growth Faults seen on the Niger Delta.
Figure 5.1.5 A) Time zero as the growth faults have not slipped. B) At a later time the growth faults have begun to slip, minimally increasing the local gradient and therefore increasing the sinuosity (dotted line shows position of river at T=0).

5.2 Magdalena

The Magdalena profile has the most complex neotectonic story (Figure 5.2.1). It is evident after examining the effects of faults intersecting a river that changes in that river (i.e. gradient and pattern) is associated with these faults. The faults intersecting the Magdalena are strike-slip faults (Figure 5.2.2). These faults are not necessarily associated with vertical movement, implying that the fault is not forcing a direct gradient change. These strike-slip faults have horizontal movement on each side of the fault relative to one another. The faults may have caused excessive fracturing of the underlying rock in such a way that accelerates erosion with the more parallel a fault intersects the river. The river slope jumps to 45-60cm/km when a fault becomes increasingly parallel with the river, also increasing the relative Concavity Index (Eq. 4) (Figures 5.2.3 and 5.2.5 both A and B).
Figure 5.2.1 Magdalena River’s Longitudinal Profile.

While the fault and river approach at 40 degrees (Figures 5.2.3 and 5.2.5 A) is more parallel then (Figures 5.2.3 and 5.2.5 B), the slope is less but shows a change from single thread channel to anastomosing. The slope and concavity reaches its maximum, 140-300 cm/km and 0.6 respectively, as the fault runs almost completely parallel to the river (Figures 5.2.3 and 5.2.5 C). These results match that of Waclawik et al. (2008) who found a marked steepening in the gradient across increasing parallel faults to a rivers flow. This steepening is then associated with the termination of alluvial overbank sediments, indicating neotectonic control on landforms (Waclawik et al., 2008). Perpendicular to the river, the strike-slip faults cause no change in slope (Figures 8-9 D). However, the intersection does show a change in channel pattern and there is a visible knickpoint just upstream of this location (Bridge and Demicco, 2008).
Examining the relative Concavity Index (Eq. 4) of Magdalena River crossing the strike-slip faults throughout the longitudinal profiles proves to be a useful measure in comparing the geomorphology of the river with respect to fault angle. The relative Concavity Index is much greater due to an increased slope, when the river runs more parallel to a fault (Figure 5.2.5). Using the four measured relative Concavity Index values, a linear...
regression with an RMS of 0.9 is fit to the data. The one outlier in the data, as discussed above, also shows the faults expression in the river as a change to anastomosing from single thread channel. The change in river pattern could thus be an explanation for the one fault’s deviation from the linear fit when comparing the angle of an intersection of a river crossing a fault to the relative Concavity Index.

Figure 5.2.3 Magdalena River crossing the strike-slip faults throughout the longitudinal profiles. Blue line is the straight line, in which concavity is measured from, with the red line the actual profile of the river coinciding with each known fault.

Bedrock lithology also has a dominant control on the behavior of a fluvial system (Bridge, 2003). The Magdalena River experiences this in the upper and lower reaches of the longitudinal profile examined (see Magdalena profile). There is a specific location in the upper reach where a decrease in slope is evident where the river approaches a confinement of flow between a resistant sandstone outcrop, and immediately thereafter experiences a
marked increase in slope (Figure 5.2.5 E). In the lower reach the Magdalena shows a marked increase in slope again coinciding with bedrock change, a transition from sandstone to limestone (Figure 5.2.5 F).

Figure 5.2.4 The Relative Concavity Index plotted against the Angle of Fault with respect to the river.

These lithic and tectonic constraints on the river system controls physical processes of the river. Specifically, changes in the river’s gradient have a control on water and sediment transport (Bridge, 2003). In regions of a slope decrease there may be permanent inundation from flooding, with increase in slope providing larger potential erosive and transport power (Bridge, 2003). Utilizing these slope measures in the 5-meander reach stretches, understanding of the inter-basin bedload transport is possible.
Figure 5.2.5 A-F illustrates the SRTM DEM (left) and Ingcominas (right) lithology map for the fault location (A-D), bedrock constriction (E), and lithology transitions (F).
Chapter 6 Bedload estimations, applying Bagnold’s Equation

The understanding of bedload fluxes, at both the global and local scales, involves identifying production sites and depositional environments (Middleton, 2003). Bedload transport is responsible for in stream habitat complexity and preservation, as well as deltaic accretion (formation) (Gomez and Church, 1989). Channel geometry and the ability of a river to recover from natural or anthropogenic disturbances, including floods and upstream impoundments are also determined by the river’s bedload transport (Gomez and Church, 1989; Knighton, 1998).

Bagnold (1966) approached bedload transport by estimating the forces needed to move an entire layer of the bed. This type of transport is termed grain flow or sheet flow (Yang, 1996). As discussed above, the delta gradient (S) is measured from the SRTM DEM along the main channel. The bedload efficiency term, (\(e_b\)) (dimensionless), is defined as the ratio of the flows capacity to do work to the amount of work done to move sediment, and depends on the fluid (air or water) in which transport occurs. The value ranges from 0.1 - 0.2 in water, with 0.1 used herein. The densities of sediment (\(\rho_s\)) and water (\(\rho\)) are 2670 kg/m\(^3\) and 1000 kg/m\(^3\) respectively. The value of 9.81 m/s\(^2\) is used for \(g\), the acceleration due to gravity. The bedload rating term (\(\beta\)) is dimensionless and, again, set to 1 for this study. \(\lambda\) is the dynamic coefficient of friction, calculated from the limiting angle of repose of sediment grains lying on the riverbed (Bagnold, 1966; Syvitski and Alcott, 1995). The limiting angle of repose at which the surface of the mass stands without shearing for most
sands is approximately 33 degrees, for which \( \tan(\lambda) \approx 0.6 \) the value used herein (Bagnold, 1966).

Applying the Bagnold equation with slope acquired from the SRTM DEM and local discharge stations, allows for understanding and analyzing inter-basin dynamics of bedload transport. Potential local sources and local sinks of the basin are obtained on a 5-meander reach basis. Only 5-meander reaches are used as this averages out local anomalous bedload transport rates by giving a value for a larger stretch of river (Schumann et al., 2008). Assuming that the stream velocity is above the critical velocity (m/s) below which no bedload transport occurs, bedload transport rates can be obtained using the local slope calculated for the Mississippi and Magdalena Rivers by utilizing local river discharge \((Q)\) stations for each of the systems (Bagnold, 1966; Nittrouer et al., 2008). Figures 10 and 11 show the potential bedload rate for the Mississippi and the Magdalena rivers respectively. This is calculated as the difference between the upstream – downstream bedload fluxes using Bagnold’s equation.

### 6.1 Mississippi

Applying the Bagnold equation to the 5 meander stretches on the Lower Mississippi provides a review of the local anomalies of the longitudinal profiles (Eq. 6). Nittrouer et al. (2008) analyzed bedload forms for the following locations: Audobon Park, Upper East Turn, and Lower East Turn. Determining a flux of \(\sim 2.2 \times 10^6\) MT/yr (70 kg/s) for the Mississippi Delta. The bedload flux for the Mississippi River’s Delta is estimated at a rate
of 59 kg/s, derived from the slope of the delta region ~1.4 cm/km from the WSE of the SRTM DEM, and using an average discharge 16,800 m$^3$/s. The 70 kg/s bedload value from Nittrouer et al. (2008) closely resembles the 59 kg/s estimated bedload.

Nittrouer et al. (2008) applied a scanner in the field and found that all bedform fluxes at the four Mississippi study sites correlated exponentially with water discharge, comparable to empirical and numerical simulations (Andrews, 1979; Kircher, 1983; Howard and Kerby, 1983; Syvitski and Alcott, 1995). Translating bedload dunes are observed at the lowest water discharges at all three (UET, LET, and Audobon Park) study sites, indicating active motion throughout the entire discharge cycle justifying the comparison of rates calculated at kg/s (Nittrouer et al., 2008).

In this study calculated potential bedload rates of the Mississippi also illustrates different zones of sources and sinks (Figure 6.1.1). At the upstream end of the Monroe Uplift there is a great potential for deposition of bedload due to the decrease in slope. Analogous to the Magdalena, the greatest depo-center for the Mississippi River appears to be the Mississippi delta. The largest potential source of bedload is that of the 5-meander reach located near the Lake County Uplift and New Madrid Seismic region (Figure 6.1.1). Here the increased slope could be attributed to an increase in water and sediment discharge of the confluence, increase in sediment flux due to seismic events, or an up-valley increase in supply rate and coarsening of grain size (due to tectonics) (Bridge, 2003).
6.2 Magdalena

The Magdalena flows through the Momposina “Mompox” tectonic depression, a low floodplain with numerous lakes encompassing an area of 800 km$^2$ (Restrepo et al., 2006). The tectonic depression traps 14%, or $\sim 20.2 \times 10^6$ MT/yr, based on estimated sedimentation rates of 2-3mm/yr settling in the massive “natural dam” (Restrepo et al., 2006a). The bedload consists of no more than 15% of the total load (IDEAM, 2001).
Restrepo and Kjerfve (2000) estimated a suspended sediment load of $144 \times 10^6$ MT/yr to the ocean, which assuming 15% of the total load, would give a bedload flux of $\sim 21.6 \times 10^6$ MT/yr ($\sim 685$ kg/s). Applying Bagnold’s bedload equation to the Magdalena River profile, reach-averaged slope is obtained for the 5-meander stretches. The river stretch flowing into the Mompox depression carries a potential bedload flux of $\sim 143$ kg/s which is decreased to 58 kg/s due to trapping in the depression. This calculated flux difference presents a loss of 85 kg/s, or 12.42%, of the Magdalena’s total bedload throughout the depression, which is approximately equal to the findings of Restrepo et al. (2006a).

The estimated potential bedload for the Magdalena illustrates zones of sources and sinks that alternate back and forth between one another (Figure 6.2.1). There appears to be three zones of potential deposition (the green segments), with the largest depo-center falling on the delta. The largest source area of bedload is the 5-meander reach just above the delta. This region is where a change in bedrock from sandstone to limestone has caused a large increase in slope. This large energy gradient has potential to erode and carry bedload just downstream of the Mompox basin, and subsequently appears to deposit the load on the delta.

Taking advantage of the high resolution of the slope, this can be used to examine bedload values found in other studies. By using values of average discharge found in Syvitski and Saito (2007), bedload fluxes are calculated and compared for eight river mouths that are found in both studies (changing only the gradient of the delta), shown in Table 6.2.1. First, the Mississippi, as discussed earlier has comparable bedload transport values for
both field calculations Nitttrouer et al. (2008) and numerical simulations Syvitski and Alcott (1995) and Syvitski and Saito (2007). Rivers in this study found to have a lower gradient, and subsequent bedload transport rate, according to measurements using the SRTM are the Eel, Fly, Indus, Magdalena, Mississippi, and Niger Deltas. The Fly Delta shows the largest difference in bedload transport, with 117 kg/s simulated according to Syvitski and Saito (2007) and 9 kg/s estimated for this study, over an order of magnitude difference. The rivers found to have a steeper gradient, and subsequent greater bedload, are the Mahanadi and Volga rivers.

Figure 6.2.1 The potential bedload, (downstream – upstream), for the Magdalena.
All of the estimated potential bedload fluxes in this study show very little differences from that of Syvitski and Saito (2007), but do show that the SRTM DEM provides a comparable data set with the ability to understand and resolve the delta slope at a higher level of detail. The high resolution SRTM DEM not only allows for a review of calculated bedload fluxes to the ocean, but also allows for understanding and analyzing inter-basin dynamics of potential bedload transport.

<table>
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<th>Dgrd* (m/m)</th>
<th>Qav* (m³/s)</th>
<th>Qb SRTM (kg/s)</th>
<th>Qb* (kg/s)</th>
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<td>41</td>
</tr>
<tr>
<td>Mississippi</td>
<td>0.00001</td>
<td>0.00002</td>
<td>15452</td>
<td>54</td>
<td>76</td>
<td>-22</td>
</tr>
<tr>
<td>Niger</td>
<td>0.00004</td>
<td>0.00007</td>
<td>6130</td>
<td>65</td>
<td>106</td>
<td>-41</td>
</tr>
<tr>
<td>Volga</td>
<td>0.00002</td>
<td>0.00002</td>
<td>8200</td>
<td>48</td>
<td>40</td>
<td>8</td>
</tr>
</tbody>
</table>

Dgrd*, Qav*, and Qb* are values taken from (Syvitski and Saito, 2007); Delta gradient, Average Discharge, and Bedload (respectively).
Chapter 7 Indus River’s Sinuosity and Lateral Migration Rates

Examining the slope vs. sinuosity plot for the entire river systems (Figure 4.1.1), it is apparent that meandering channels are an important end-member in river planform classifications and is a common pattern found in nature (Hickin, 2003). The planform geometry of meandering rivers is controlled by lateral migration processes, in which the outer banks of a channel’s bends erodes matched by the relative deposition rate of the inner point bar on the bend (Knighton, 1998; Hickin, 2003). The sediment available for point bar deposition is related to the sediments deposited forming the meandering river’s floodplains through lateral migration (Hickin, 2003). The process of lateral migration is one of the two main processes responsible for development of sand bodies, and as stated before, is of interest in the field of Sedimentary Petrology (Richardson et al., 1987).

Lateral migration can be determined by comparing two or more time intervals of a river’s centerline (Knighton, 1998; Hickin, 2003). This will provide a lateral migration rate over the time-span analyzed. Starting with the centerline for the earliest date of the two compared times, a perpendicular line (to the river’s centerline) is created every two kilometers (Figure 7.1 A). Each of the perpendicular lines are clipped by using the two centerline locations of the river as boundaries, leaving line segments that indicate the distance between the times compared (Figure 7.1 B). The average of these line segment lengths is the average of the distance the center of the river has moved, via lateral migration, over the time span between the two time steps.
Figure 7.1 reflects the two essential steps (A. creation of, and B. clipping of perpendicular lines) in determining lateral migration rates.

7.1 Indus River

Using the Indus River as a test case, sinuosity and lateral migration rates are compared for the entire Indus River’s lower longitudinal profile examined, as well as the River’s fluvial and tidal reaches (Figure 7.1.1). The time stamps used in this study are the years 1944 (Toposheets), 2001 (SRTM), and 2010 (MODIS) both pre and post-flood. The 1944 maps is at a scale of 1:63,000 to 1:250,000 and of comparable resolution to the 90 m SRTM maps. They are compiled mainly from Survey of India Maps, with aerial photo updates prepared by Army Map Service (GPDE), Corps of Engineers, U.S. Army Washington DC (See list of Maps used at the end of the chapter).

The year 1944 is chosen as the start year for this analyses as this is before the large-scale reduction of water and sediment discharge (after the 1950s) due to the emplacement of
barrages and dams (more than 70% for water and 80% for sediment; Milliman et al., 1984). The study concludes with the Indus River flooding that occurred for two months starting in July of 2010, which inundated more than 35,000 km$^2$ and displacing ≥3 million Pakistani citizens (Syvitski et al., 2010). The flood, exceeding 30,000 m$^3$/s, indicates that traditional river gauging is obsolete and requires the use of satellite data to understand the ever-changing fluvial domain (Syvitski et al., 2010).

Comparing the differences over the years reveals a relatively narrow bound of values for the sinuosity of the Indus River’s lower longitudinal profile (Table 7.1.1). The Indus’ sinuosity provided by the 1944 maps reveals a meandering system with a sinuosity of 1.61. The sinuosity increases by 28% to a value of 1.78 over a span of 56 years to the year 2000, where after the river sinuosity slightly increases to 1.81 by 2010, just before the Indus flood in July. The sinuosity of the Indus River then takes a dramatic 12% decrease to 1.71 following the large flood of 2010. The Indus, as a whole, appears to slowly allow its meanders to migrate before the large flood event lowers the Indus’ sinuosity precisely between maximum and minimum calculated over a 66-year time span.

The lateral migration of the modern Indus is calculated every 2 km between the river centerlines at the specified intervals (Table 7.1.2). The lateral migration between the first two river centerlines is on average 1.95 km (as measured with an accuracy of 90 m) for the 56 years between 1944 and 2000 or 35 m/y. The next timespan compared, between the 2000 SRTM survey and the 2010 MODIS pre-flood imagery, resulted in a mean lateral migration of 124.9 m or 11 m/y. Between the last two dates, just days before the flood
Figure 7.1.1 A) The Indus’ lower floodplain and delta. From A to C (black letters) the entire longitudinal profile examined was analyzed for sinuosity and lateral migration rates. B) Inset map of the upper floodplain in which the fluvial component is measured for sinuosity and lateral accretion. The outline box of D is where the Figure 1.1 was taken. C) Shows the delta where the tidal effects were analyzed between B (Thatta) and the furthest most downstream merging of all time spans.
and just after the flood in 2010, the lateral migration rate is 338.9 m per 52 days as measured with MODIS imagery. The average lateral migration measurement between the 2000 SRTM imagery and the post-flood imagery, is 463.8 or 42 m/y a value close to the historical value of 35 m/y.

This sinuosity and lateral migration calculations are examined at a smaller scale, in an attempt to separate fluvial and tidal influence, as the limit of tidal effects upstream are known. With tidal effects felt upstream as far as Thatta, this will serve as the breaking point in which to measure the sinuosity and lateral migration rates for the tidally influenced portion below Thatta and the fluvial dominated region above (Figure 7.1.1; Eisma and Lubberts, 1998). The sinuosity of the Indus River’s fluvial stretch in 1944, 2000, 2010 pre and post-flood the values is 1.63, 1.81, 1.82, 1.71 respectively. The sinuosity increased 30% on average between 1944 and 2000 where after the Indus experiences a small increase, just 1%, over the next ten years. The sinuosity then decreased 13%, following the Indus flood, but not to a value below the 1.6 threshold, established in the previous chapter, for a meandering system.

The tidal region shows a similar transition for sinuosity, but the rate of change is distributed at a steadier rate of increase over the first two comparison timespans. The Toposheet 1944 date revealed a low sinuosity of 1.48, with an increase to 1.65 calculated for the SRTM in the year 2000 (an increase 35%). Throughout the next 10 years, by 2010 the sinuosity had increased by 16% to 1.75 as found from MODIS. While the change in sinuosity halved between the time spans measured, it did not deviate between the three earliest comparison dates as the upper portion of the river, above the tidal reach, showed.
The flood did however reduce the sinuosity by 7%, appearing to represent a trend in this system to straighten the river in an attempt to accommodate the large flow of the flood.

In examining the lateral migration between the first two river centerlines of the separated fluvial and tidal stretches for the 56 years between 1944 and 2000 are on average 2.03 km or 36 m/y, and 1.65 km or 30 m/y respectively. The comparison between 2000 (SRTM) and 2010 (MODIS pre-flood) imagery, shows an migration of 152 m for the fluvial stretch and only 56.6 m for the tidal stretch, or 14 m/y and 5 m/y respectively. Between the last two dates, just days before the flood and just after the flood in 2010 the lateral migration is 372.3 m for the fluvial stretch and 198.5 m for the tidal stretch per 52 days as measured with MODIS imagery. The average lateral migration rate measured between the 2000 SRTM imagery and the post-flood imagery is on average 524.3 m for the fluvial stretch and 42.6 m for the tidal stretch, or 48 m/y and 4 m/y respectively.

### Table 7.1.1 Planform Sinuosity of the Indus River

<table>
<thead>
<tr>
<th>Date/Dataset</th>
<th>Sinuosity of Entire Lower Profile</th>
<th>Sinuosity Fluvial Reach</th>
<th>Sinuosity Tidal Reach</th>
</tr>
</thead>
<tbody>
<tr>
<td>Toposheet 1944</td>
<td>1.61</td>
<td>1.63</td>
<td>1.48</td>
</tr>
<tr>
<td>SRTM 2000</td>
<td>1.78</td>
<td>1.81</td>
<td>1.65</td>
</tr>
<tr>
<td>Pre-Flood 2010</td>
<td>1.81</td>
<td>1.82</td>
<td>1.75</td>
</tr>
<tr>
<td>Post-Flood 2010</td>
<td>1.71</td>
<td>1.71</td>
<td>1.70</td>
</tr>
</tbody>
</table>
Table 7.1.2 Lateral Migration Rates of the Indus River

<table>
<thead>
<tr>
<th>Time-Span</th>
<th>Lateral Migration Entire Lower Profile (m/y)</th>
<th>Lateral Migration Fluvial Reach</th>
<th>Lateral Migration Tidal Reach</th>
</tr>
</thead>
<tbody>
<tr>
<td>1944-2000</td>
<td>35</td>
<td>36</td>
<td>30</td>
</tr>
<tr>
<td>2000-2010(pre)</td>
<td>11</td>
<td>14</td>
<td>5</td>
</tr>
<tr>
<td>2000-2010(post)</td>
<td>42</td>
<td>48</td>
<td>4</td>
</tr>
</tbody>
</table>

It is evident that the sinuosity reveals a relatively narrow bound of values for the Indus River’s lower longitudinal profile. Breaking apart the fluvial and tidal reach of the Indus River’s lower longitudinal profile exposes that the fluvial reach is on average 22% more sinuous than the tidal reach, except for the values following the 2010 flood. The tidal reach lateral migration rates are the lowest, and show a marked drop following the year 2000. This could be due to tidal channels experiencing stability rather than migration as found in other tidal deltas with a large tidal component (Wolanski and Eagle, 1991). The loss of the Indus River’s water and sediment discharge to the system could be causing the channel’s stability found in the mouth. The lateral migration of the fluvial reach is about equivalent to the entire Indus River’s lower longitudinal profile as it makes up over 70% of the length of the profile, and appears to be the dominant zone of migration. With the lateral migration values between 2010 post-flood and 2000 closely matching the historical rate, it can be concluded that meander movement is not only a steady state process as largely modeled, but is greatly influenced by short duration, low frequency, high-intensity events that speed meander migration and create oxbow-lakes through channel cutoffs.

This single flood event illustrates the dynamic behavior of a fluvial system and why there is a need to develop predictive relationships on controlling processes for floodplain...
Maps

- NF42-02, Series U502, Title Lakphat, Pakistan & India, Compiled in 1955 from one-inch series, 1:63,360 Survey of India, 1883-1929, with planimetric detail revisions from 1944 aerial photography
- NG42-13, Series U502, Title Karachi, Pakistan, Compiled in 1954 from one-inch series, 1:63,360 Survey of India, 1937-1938, with planimetric detail revisions from 1944 aerial photography.
- NG42-06, Series U502, Title Nawabshah, Pakistan, Compiled in 1954 from Quarter-inch series 1:253, 440, Survey of India, 1942, with planimetric detail revisions from 1944 aerial photography.
Chapter 8 Oxbow Lakes in the Fly and Mississippi River Floodplains

Lowland river channels wind through their valley forming meander patterns. Growth and cutoff of meanders are important in the construction of the meandering river’s floodplain (Wolman and Leopold, 1957), and its alluvial architecture (Allen, 1965: Bridge et al., 1986; Constantine and Dunne, 2008). Predictive relationships for oxbow lake formation and geometry are of great interest to the hydrocarbon industry due to these bodies’ ability to capture and sequester hydrocarbons (Richardson et al., 1987).

Oxbow lakes form by the process of meander cutoff, when the main channel reroutes its main course favoring a straighter, higher gradient, pathway (Figure 8.1). Cutoffs are typically initiated or accelerated during high discharge events, and their occurrence influences rates of meander migration (Hudson and Kesel, 2000; Constantine and Dunne, 2008). Subsequently, the abandoned channels fill with silt and clay during floods only and these fine sediments can form effective barriers to lateral flow (Richardson et al., 1987). A typical cutoff period is relatively brief taking only ~2-10 yr (Gagliano and Howard, 1984). Infill and sedimentation rates drop off sharply as the distance from the river increases, with coarse sand grains making up the abandoned channel bed and, crevasse splays, peat, silt, and clay capping the sand bodies after abandonment (Gagliano and Howard, 1984; Richardson et al., 1987). The process terrestrialization refers to the infill, and eventual growth of vegetation in the oxbow lake (Piégay et al. 2001; Figures 8.1.1 and 8.2.1).
In our overarching study of topographical characteristics of major river systems, we classified seven river systems as meandering rivers, since their main channels have an average sinuosity >1.6 along their entire longitudinal profile (Figure 4.1.1 and Figure 8.2). All these systems are actively migrating, creating complex point bars and abandoning oxbow lakes throughout their floodplains.

This study examines two selected major fluvial systems, distinctly different in their drainage basin size, climate zone, water and sediment discharges. The two systems are chosen for two reasons: 1. They are well-researched systems, with abundant literature on floodplain analysis and 2. They represent end members of the seven meandering rivers (Figure 8.2). However, the floodplains of the Fly River, Papua New Guinea and the continental-scale Mississippi River in North America, have one thing in common; these fluvial systems are both meandering rivers with many oxbow lakes populating their floodplains. The topographic data analysis uses Shuttle Radar Topography Mission Digital Elevation Model data, SRTM, in combination with the Surface Water Body Dataset,
SWBD, to study the abundance of oxbow lakes, the occurrence of oxbow lakes with respect to channel slope and sinuosity, and the characteristics of oxbow lakes with distance from the active channel belt.

Figure 8.2 Slope-sinuosity relationships along the longitudinal profile 0-100 m elevation for meandering systems based on SRTM data analysis. The Fly River and the Mississippi River form the end-members of the selected meandering rivers.

Figure 8.3 Location map of floodplains of A) the Fly River and B) the Mississippi River. The black line represent the river’s centerline, the blue zone represents the selected stretches containing the many oxbow lakes.
8.1 Middle Fly River

The Fly River in Papua New Guinea has a drainage basin size over $61 \times 10^3$ km$^2$ and a mean annual water discharge of 6,500 m$^3$/s, with a peak discharge $\sim$7,000 m$^3$/s (Syvitski et al., 2005). The SRTM data analysis shows that the average channel slope for the lower Fly for the last $+800$ km is 4 cm/km and the active channel has a channel sinuosity of 2.3. The riverbed is dominated by sand, and the river carries an average suspended sediment concentration of 100 mg/l in addition (measured between 1950 and 1991) (Mossa, 1996; Rowland et al., 2005).

To analyze oxbow lakes abundance and geometry, we focus on the Middle Fly River where the tidal influences are mostly negligible. The Middle Fly River, where the majority of the oxbow lakes are found, only drains an area of $18 \times 10^3$ km$^2$ and has a mean annual water discharge of 2,240 m$^3$/s (Ok Tedi Mining Ltd, 1988; Dietrich et al., 1999). The slope of the Middle Fly River is 3.6 cm/km and the active channel locally has a sinuosity of 2.6. The container valley includes both an inner and outer floodplain, ranging in width from 4 to 14 km (Salomons and Eagle, 1990). The innermost floodplain of $\sim$2 km width contains scroll-bar complexes, old alluvial deposits, and oxbow lakes (Salomons and Eagle, 1990; Dietrich et al., 1999). The outer floodplain is composed of blocked valley lakes and backswamps (Salomons and Eagle, 1990; Dietrich et al., 1999).

The Middle Fly River transitions from rainforest dominated floodplain to a swamp grass system in its lower reaches (Dietrich et al., 1999). The Strickland River creates a backwater on the Fly River, where they converge, with this confluence associated with
swamp formation and reduction of channel migration rates (Dietrich et al., 1999). The basin receives \(\sim 10\) m/yr of rainfall in the uplands, with the lowlands still receiving in excess of 2 m/yr (Dietrich et al., 1999). The lower swamp grass dominated region experiences rain-induced flooding yearly (Dietrich et al., 1999).

![Image of open water and infilled oxbow lakes](image)

Figure 8.1.1 A March 2, 2011 Terrametrics image shows open water and infilled oxbow lakes. The oxbow lake at the bottom right of the image shows an oxbow still receiving sediment from the main Fly River channel.

Humans have largely influenced the Fly River system since 1985, with the introduction of the Ok Tedi mine. The Ok Tedi mine has introduced 1720 million tons of sediment, rock waste and tailings, between 1985-2010 (Ok Tedi Mining Limited, 1988). This mine has caused the Middle Fly River to experience a four fold increase in its suspended sediment load (Rowland et al., 2005).

The sediment load of the pristine Fly River downstream at its delta, is estimated to be 80
to 90 million MT/yr, with only 30% of that load coarser than silt (Dietrich et al., 1999). Clay content in fine-grained deposits reaches a maximum of \( \sim 20\% \), leaving the last 50% of the load silt (Rowland et al., 2005). The sedimentation along the Fly River kept pace with Holocene sea level rise; this sedimentation led to blocking of lowland tributaries and lake formation, hence the large blocked valley lakes seen in Figure 8.2.2 (Blake and Ollier; 1971). Pickup (1984) estimated Holocene floodplain sedimentation rates, along the Middle Fly, to be \( \sim 0.1 \) cm/yr. Vertical accretion rates, taken by Rowland et al. (2005), range between \( 0.25 \pm 0.04 \) cm/yr and \( 0.7 \pm 0.09 \) cm/yr. Dietrich et al. (1999) established an infill rate for oxbow lakes to be at least \( \sim 900 \) years.

### 8.2 Lower Mississippi River

The Mississippi River has a drainage basin size over \( 3.2 \times 10^6 \) km\(^2\) and a mean annual water discharge of 15,450 m\(^3\)/s, and a peak discharge of 30,480 m\(^3\)/s (Syvitski et al., 2005). The average slope for the lower Mississippi floodplains for the last +1600 km is 6 cm/km and this stretch has an average channel sinuosity of 1.7 (Figure 8.2). The riverbed is dominated by fine sand, and the river carries an average suspended sediment concentration of 420 mg/l (measured between 1950 and 1991) (Mossa, 1996; Rowland et al., 2005).

The container valley of the Lower Mississippi River ranges in width from 30 to 100 km (Aslan and Autin, 1999, Figure 8.3). The container valley contains terraced, braided river deposits of Pleistocene age, five Holocene Mississippi River meander belts, Holocene back swamps and has a lower deltaic plain out to the Mississippi bird foot delta (Aslan and Autin, 1999; Saucier and Snead 1989; McGinley, 2009). In the southern Lower
Mississippi Valley, Mississippi River meander belts are 5–15 km wide and consist of sinuous scroll bars (Aslan and Autin, 1999). This southern portion up to ~300 km before the mouth of the river, extending a distance of ~800 km upstream, is where the oxbow lakes are most abundant. The channel slope of the Lower Mississippi is 6.3 cm/km and the active channel has a sinuosity of 1.6 as calculated from the SRTM topographical data.

Precipitation in this region varies between 1,100 mm/yr further north near Cairo, Illinois and 1,400 mm/yr near Natchez, Mississippi (McGinley, 2009). The Mississippi experiences yearly spring flood waves, with records indicating 16 years in which major flooding occurred on the Mississippi since 1927 (Smith et al., 1990). Before widespread agriculture dominated, the floodplain vegetation consisted of deciduous forest and swamp vine (McGinley, 2009).

Humans are profoundly influencing the lower Mississippi River system. Concrete bank revetment has reduced bank erosion potential, limiting lateral movement of the main channel (Harmar and Clifford, 2007). Engineering intervention performed on the Lower Mississippi River’s profile artificially shortened the river by 230 km, straightening 14 of the most sinuous bends between Old River and Memphis between 1930-43 (Harmar and Clifford, 2007; Moore, 1972; Winkley, 1977). There has also been extensive dredging and gravel mining, decreasing the amount of coarse material in the Lower Mississippi River (Harmar and Clifford, 2007). The system has also experienced dramatic increases and decreases of sediment load due to a period of poor farming practices and implementation of good farming practices (Kesel, 1988; Harmar and Clifford, 2007).
The sediment load of the Mississippi is 200 million MT/yr, with only 10% of that load coarser than silt (Meade, 1995; Mossa, 1996). The silt content of the Mississippi is ~70 percent, with the clay making up the rest of the suspended load (Mossa, 1996). Vertical floodplain accretion rates, taken by Rowland et al. (2005), range between 4.5±0.6 cm/yr and 5.8±1.5 cm/yr. A study of oxbow lakes shows ages of oxbow lakes can reach 3-5000 years old (Wren et al., 2008). Literature research has found no established rates of infill for the Mississippi. However, Hudson et al. (2008) performed a study in which they examined 25 oxbow lakes varying in ages of cutoff. Using his data a range of infill rates can be established. The sixteen lakes examined in the study by Hudson et al. (2008),
found within the region of embankment along the Mississippi, have infilled on average ~68% over the last ~73 years. Extrapolating this rate out linearly would yield an infill rate of only ~110 years. The use of a linear infill rate probably results in an underestimation of infill duration, since it is likely that sedimentation rates decrease over time, as the main channel migrates away.

Table 8.2.1 Fluvial and Floodplain Characteristics of the Mississippi and Fly Rivers.

<table>
<thead>
<tr>
<th></th>
<th>Units</th>
<th>Mississippi</th>
<th>Fly</th>
</tr>
</thead>
<tbody>
<tr>
<td>Length of the studied river reach</td>
<td>km</td>
<td>800</td>
<td>160</td>
</tr>
<tr>
<td>Meander belt width</td>
<td>km</td>
<td>15</td>
<td>2</td>
</tr>
<tr>
<td>Total floodplain Area Examined</td>
<td>km²</td>
<td>15,000</td>
<td>640</td>
</tr>
<tr>
<td>Climatic regime</td>
<td></td>
<td>Temperate</td>
<td>Tropical</td>
</tr>
<tr>
<td>Vegetation</td>
<td></td>
<td>Deciduous, Swamp</td>
<td>Rain Forest</td>
</tr>
<tr>
<td>Human impact, # of Interventions</td>
<td>-</td>
<td>3</td>
<td>1</td>
</tr>
<tr>
<td>Drainage area approaching selected river stretch</td>
<td>km²</td>
<td>3,200,000</td>
<td>18,000</td>
</tr>
<tr>
<td>Mean annual water discharge</td>
<td>m³/s</td>
<td>15,450</td>
<td>2,240</td>
</tr>
<tr>
<td>Peak annual water discharge</td>
<td>m³/s</td>
<td>30,480</td>
<td>7,000</td>
</tr>
<tr>
<td>Sediment load</td>
<td>10⁶ MT/yr</td>
<td>200</td>
<td>80</td>
</tr>
<tr>
<td>Flooding recurrence Interval</td>
<td>yr</td>
<td>5</td>
<td>1</td>
</tr>
<tr>
<td>Mean channel slope</td>
<td>cm/km</td>
<td>6.3</td>
<td>3.6</td>
</tr>
<tr>
<td>Mean channel planform sinuosity</td>
<td>-</td>
<td>1.6</td>
<td>2.6</td>
</tr>
<tr>
<td>Lateral migration rates*</td>
<td>m/yr</td>
<td>50</td>
<td>2</td>
</tr>
<tr>
<td>Floodplain aggradation rate</td>
<td>cm/yr</td>
<td>5</td>
<td>0.5</td>
</tr>
<tr>
<td>Oxbow lake Max Ages</td>
<td>yr</td>
<td>300</td>
<td>900</td>
</tr>
</tbody>
</table>

* Hudson and Kesel (2000); Dietrich et al. (1999)

Oxbow lakes are hand-selected from a combination of the SRTM DEM and the water bodies in the SWBD (Surface Water Body Dataset). The minimum lake size delineated for the SWBD are features comprised of a length > 600m and width >183m. The lakes are confirmed to be oxbow lakes by their curved shape and/or location within a scroll bar complex. The oxbow lake area is calculated for each individual oxbow lake based on the
projection of the data into their corresponding UTM projection. Channel sinuosity and channel slope are calculated along the length of the entire longitudinal profile in the selected zones with abundant oxbow lakes and at distinct 5-meander length reaches of the channel to allow investigation of these parameters in relation to the oxbow geometry and abundance.

The river centerline, which is extracted from the SRTM DEM, is utilized to create stepwise increasing buffer zones with respect to the river's centerline out to the limit of the active floodplain (Figure 8.2.2). The outer limits of the active floodplain are determined from field studies reported in literature. The buffer zone distances step up by 100m intervals, tallying the ones with a change in cumulative oxbow lake area up to 2 km for the Fly River floodplain. The buffer zone distance step up by 1 km intervals, again tallying only buffers experiencing a positive change in cumulative oxbow lake area, to 15 km for the Mississippi River floodplain. Using the buffer zones, individual oxbow lakes are classified to be located within buffer distances from the river's centerline. It is important to note that the buffer size is in either direction perpendicular to the river, so the 1 km buffer zone extends 1 km on either side of the river. Subsequently, the total cumulative area for all selected oxbow lakes within each additional buffer zone from the main channel is calculated. With the buffers established to the outer edge of the meander scroll bar complex, a cumulative oxbow lake area vs. distance from river centerline relationship can be determined.
8.3 Middle Fly River Results

A 640-km² area of floodplain (4 km wide floodplain * 160 km reach) is examined for the Middle Fly River. In this buffer zone 48 separate oxbow lakes are identified. The smallest lake found in this study had an area of ~0.01 km², the largest ~8.4 km² with an average area of 0.95 ±1.4 km². The total oxbow lake area is 45.8 km², or ~7% of the floodplain examined. The frequency distribution of oxbow lake area in the Fly River’s floodplain showed that the smallest lakes, 0 -0.5 km², are the most common. The largest lakes occur very infrequently (Figure 8.3.1).
Figure 8.3.1 Distribution of oxbow lake area in the Fly River floodplain.

There is an interesting pattern in the average oxbow lake area with increasing distance away from the active Fly River channel. There is a weak trend for the average oxbow lake area to decrease away from the Fly River (Figure 8.3.2). However, there is a dramatic drop in the average area of oxbow lakes for the buffers created at 3-700 meter distances. This could be due to rapid infill with both sediment and vegetation of oxbow lakes within the 700-meter distance of the active channel since the increase in sediment load from the Ok Tedi Mine (Hudson et al. 2008; Day et al. 2008).
Figure 8.3.2 Mean oxbow lake area measuring as a function of buffer zone distance away from the Fly River’s active channel.

Figure 8.3.3 Sinuosity of the 5-meander stretches plotted with total oxbow lake area.
It is expected to see a decreasing oxbow lake area with increasing channel sinuosity, arguing that if cutoffs had generated oxbow lakes recently, the channel would appear to be straighter. In lowermost stretches of the studied Middle Fly River, tidal currents and backwater effects enlarge the channel width. It is decided to exclude some of the anomalously large oxbow lakes that occur in this specific stretch from the analysis. By removing the 5-meander reaches influenced by the tide, a weak trend of decreasing total lake area is seen with a greater sinuosity (Figure 8.3.3).

\[ A_{OL} = 5.7S_i^{-1.1} \]  
Eq. 7

\( A_{OL} \) cumulative oxbow lake area in km\(^2\)

\( S_i \) channel sinuosity

The regression analysis of cumulative oxbow lake area with distance from the main channel of the Fly River revealed a logarithmic relationship with a strong correlation (R\(^2\) ~0.95) (Figure 8.3.4).

\[ A_{OL} = 18.7 \ln D + 35.6 \]  
Eq. 8

\( A_{OL} \) cumulative oxbow lake area in km\(^2\)

\( D \) distance from active river channel in km

The furthest distance from the Fly River that oxbow lakes are found is ~2km out from the river’s centerline. The largest deviation in the regression is between the 700 and 800 m buffers, in which an 8.4 km\(^2\) oxbow lake is found in reach of the Fly River system that is affected by tidal currents and backwater. Utilizing the maximum lateral migration rate, calculated by Dietrich et al. (1999) of 2 m/yr, it would take ~1000 years to rework the
floodplain in either direction. The infill rate of at least ~900 years established by Dietrich et al. (1999) matches the time it would take for the Fly River to rework the scroll complex based on the maximum lateral migration rate, 2 m/yr (Figure 8.3.4 and Table 8.2.1). By using the lateral migration rate of ~1 m/yr taken by Rowland et al. (2005), along with the minimum rate established by Dietrich et al. (1999), it would take ~2000 years (or ~1000 years) in either direction for the floodplain to be reworked by the river. It is interesting to see that the infill rate and the reworking rate are synonymous with one another.

Another intriguing find is the floodplain sedimentation rate. Day et al. (2008) found that the sediment carried during floods deposits rapidly, with the sediment deposition reaching a maximum of 1 km on either side. Examining the cumulative oxbow lake area within the 1 km buffer distance around the Fly River centerline reveals that ~90 percent (or 40/45 km$^2$) of the total oxbow lake area exists within this 2 km wide region on both sides of the main channel (Figure 8.3.4). These oxbow lakes must be relatively young features, as they sequester the majority of the overbank sedimentation and are still in existence.
Figure 8.3.4 Cumulative oxbow lake area, against distance, from the centerline of the active Fly River. Note the large increase in oxbow lake area between the 700m and 800m buffer, which is due to an 8.4 km$^2$ oxbow lake found in the tidal reach of the system. (Formation year based on a 2 m/yr lateral migration rate).

### 8.4 Lower Mississippi River Results

A 15,000-km$^2$ area of floodplain (30 km wide floodplain * 500 km reach) is examined for the Lower Mississippi. In this buffer zone 182 separate oxbow lakes are identified. The smallest oxbow lake found in this study had an area of ~0.02 km$^2$, the largest ~18.4 km$^2$ with an average area of 1.9 ±3.3 km$^2$. The total oxbow lake area is 349.1 km$^2$, or ~2% of the floodplain area examined. The frequency distribution of oxbow lake area in the Mississippi River’s floodplain showed that the smallest lakes, 0 -0.5 km$^2$, where the most common and the largest lakes, again, very infrequent (Figure 8.4.1).

There is a strong trend for the average oxbow lake area to decrease away from the Mississippi River (Figure 8.4.2). Another strong relationship shows that a decreasing
total lake area is established with a greater main channel sinuosity (Figure 8.4.3).

\[ A_{OL} = 317.5 S_i^{-7.2} \quad \text{Eq. 9} \]

\( A_{OL} \) cumulative oxbow lake area in km\(^2\)

\( S_i \) channel sinuosity

![Distribution of Oxbow Lake Size: Mississippi](image)

Figure 8.4.1 Distribution of oxbow lake area in the Mississippi River floodplain.

Prehistoric oxbow lakes used in the study by Hudson et al. (2008): Lake Bruin, Lake Chicot, Lake Providence, and Lake Saint John have infilled by 20%, 21%, 47%, and 52%, respectively. These prehistoric oxbow lakes have no year associated with time of cutoff, because there are no maps in this area before their time of cutoff. The other five oxbow lakes, found outside the embankment, have infilled on average ~42% over the last ~152 years. Extrapolating this rate out at a linear rate would give an infill rate of ~370 years. Again sedimentation drops off significantly as the river migrates away from the cutoff, so this would be toward a minimum rate.
Figure 8.4.2 Mean oxbow lake area from the buffer zone distance away from the Mississippi River’s main channel.

Figure 8.4.3 Sinuosity of the 5-meander stretches plotted with total oxbow lake area.
The regression analysis of cumulative oxbow lake area away from the Mississippi River’s main channel revealed another strong correlation with a $R^2 \sim 0.92$ (Figure 8.4.4).

$$A_{OL} = 110.8 \ln D + 86.0$$  \hspace{1cm} \text{Eq. 10}

$A_{OL}$ cumulative oxbow lake area in km$^2$

$D$ distance from active river channel in km

Oxbow lakes are found at maximum distance of ~15km out from the active Mississippi’s River channel. Hudson and Kesel (2000) established migration of meander bends averaging 45.2 m/yr in the upper alluvial valley of the Mississippi, where there are numerous clay plugs, and an increase to 59.1 m/yr in the lower alluvial valley, where there are fewer clay plugs. By assuming an average migration rate of 50 m/yr, as the river’s centerline is not split into upper and lower reaches, this provides ~300 yr time span in either direction for the Mississippi to rework the meander complex (Figure 8.4.4 and Table 8.2.1). Using the lower and upper limit on migration rates established by Hudson and Kesel (2000) would provide a reworking age of ~250-330 years. The reworking age would be the time it would take for the river to traverse the floodplain in one direction. The infill age and the reworking age, again like the Fly, are approximately equal to each other, assuming the rates calculated are correct. Examining the cumulative oxbow lake area within the 5 km buffer distance around the Mississippi River centerline reveals that ~86 percent (or ~300/350 km$^2$) of the total oxbow lake area exists within this 10 km region (Figure 8.4.4).
The Fly River and the Mississippi River are selected as end-members of this studied set of meandering rivers. The Fly River shows the largest sinuosity, ~2.3, of all systems, and the Mississippi River has a more modest sinuosity of ~1.7. The SRTM data of the floodplains of both river systems show abundant scroll bar complexes and oxbow lakes. The oxbow lake area is calculated for each individual oxbow lake based on the projecting the data into their corresponding UTM projection. Channel sinuosity and channel slope are calculated along the length of river with abundant oxbow lakes and at distinct 5-meander length reaches. This enabled investigation of these parameters in relation to the oxbow geometry and abundance. Buffer zones are established based on floodplain widths of the system to identify the distance of an oxbow lake from the active channel.
Subsequently, the total cumulative area for all selected oxbow lakes within each additional buffer zone from the main channel is calculated establishing a cumulative oxbow lake area vs. distance relative to the active channel.

The Fly and Mississippi floodplain areas consist of 7% and 2% oxbow lake area respectively. It is important to note that the calculations used depend on the chosen value for the outer boundary of the active floodplain, and the Mississippi meander belt width can locally be much narrower than the 15 km used in these presented results. Using the smallest meander belt width as defined by Aslan and Autin (1999), 5 km, would cause the relative percentage of oxbow lake area to increase to ~7% which is similar to the relative oxbow lake area found in the Fly River floodplains.

The frequency distribution of oxbow lake area in both floodplain systems showed that the smallest lakes, 0 -0.5 km², where the most common (45-50%) and the largest lakes occur infrequently (<5%; Figures 8.3.1 and 8.4.1). There is a weak trend for the average oxbow lake area to decrease away from the Fly River (Figure 8.3.2), and a much stronger trend for the Mississippi River (8.4.2). The most proximal 50% of the Fly River floodplain area, relative to the main channel, harbors 90% of the oxbow lake area. The nearest 67% of the Mississippi River floodplain area, relative to the main channel, holds 86% of the oxbow lake area.

In the lowermost stretches of the studied Middle Fly River, tidal currents and backwater effects enlarge the channel width. It is decided to exclude some of the anomalously large
oxbow lakes that occur in this specific stretch from the analysis. By removing the 5-
meander reaches for the Fly River that are influenced by tidal processes, a trend of
decreasing total lake area with a higher channel sinuosity is apparent (Figure 8.3.3). The
negative correlation between channel sinuosity and oxbow lake area is much stronger for
the Mississippi River, showing that a more sinuous stretch could be on the verge of
cutting off a bend and producing an oxbow lake due to the lack of oxbow lake area in the
vicinity (Figure 8.4.3).

Table 8.4.2 Oxbow Lake Stats and Equation Parameters for the Mississippi and Fly Rivers

<table>
<thead>
<tr>
<th></th>
<th>Units</th>
<th>Mississippi</th>
<th>Fly</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean Oxbow lake area</td>
<td>km²</td>
<td>1.9</td>
<td>0.95</td>
</tr>
<tr>
<td>Max Oxbow lake area</td>
<td>km²</td>
<td>18.4</td>
<td>8.4</td>
</tr>
<tr>
<td>StDev. Oxbow lake area</td>
<td>km²</td>
<td>3.3</td>
<td>1.4</td>
</tr>
<tr>
<td>Lake Area vs. Distance Coefficient</td>
<td>-</td>
<td>110.8</td>
<td>18.17</td>
</tr>
<tr>
<td>Lake Area vs. Dist. X-int Vertical Shift</td>
<td>-</td>
<td>86</td>
<td>35.6</td>
</tr>
<tr>
<td>Lake Area vs. Sinuosity Coefficient</td>
<td>-</td>
<td>317.5</td>
<td>5.7</td>
</tr>
<tr>
<td>Lake Area vs. Sinuosity Exponent</td>
<td>-</td>
<td>-7.2</td>
<td>-1.1</td>
</tr>
</tbody>
</table>

Regression analysis yielded a strong relationship for cumulative oxbow lake area with
distance from the river’s main channel in each fluvial basin. The Fly River and oxbow
lake analysis yielded the equation: \( A_{OL} = 18.7 \ln D + 35.6 \) with a high correlation (\( R^2 \) of
0.95). The Mississippi River and oxbow lake analysis yielded the equation:
\( A_{OL} = 110.8 \ln D + 86.0 \) with a high correlation (\( R^2 \) of 0.92). These results can be utilized to
improve river meandering models and illuminate the physical processes by which a river
populates a floodplain with oxbow lakes through channel cutoff.

There are several interesting scaling relationships found between the two fluvial systems
and their oxbow lake geometries (Tables 8.2.1 and 8.4.2). The lake area vs. distance
coefficient and lake area vs. sinuosity exponent scale with both the meander belt width and mean annual discharge (ratio ~0.15). This can be explained because the meander belt width constrains the distance a lake can be from the main channel, and mean annual discharge has a degree of control on both the channel’s sinuosity and width (oxbow lake area). The oxbow lake area distribution’s primary statistical characteristics i.e. mean, max, standard deviation, and lake area vs. distance vertical shift scale with channel slope and sediment load (~0.4-0.5). This could potentially be explained due to the fact that the channel slope and sediment load can constrain the size of oxbow lakes through rate of cutoff (gradient) and infill rate (sediment load).

Another parallel drawn between the two systems is that the infill rate and migration/reworking rate illustrate that there is a ceiling on the uppermost age of an oxbow lake before it is either infilled or reworked. This illustrates why previous studies have hinted that assessing both the vertical and lateral movement of the main river channel, over time, can help determine the sedimentation rates and lifespan of former channels (oxbow lakes) (Gagliano and Howard, 1983; Piégay et al. 2001).

This study shows that SRTM topographical data for meandering river systems can aid in understanding the process of lateral migration and infill rates. The Fly and Mississippi Rivers have both been influence heavily by human intervention (Rowland et al., 2005; Kesel, 1988). The Middle Fly River has seen a 4 fold increase in its suspended sediment load following post-mine operations (Rowland et al., 2005). Data collected along the Lower Mississippi suggest that an overall decrease in average annual suspended load...
since 1850 exceeds 70% (Kesel, 1988; Meade and Moody, 2010).

The relationships that can be drawn from the SRTM data will help expand the knowledge of how sediment can be sequestered in these systems. Rowland and Dietrich (2006), upon examining the sediment volume sequestered in an oxbow lake (through a tie-channel) along the Mississippi, found that between 1851 and 1998 24% of the lake had been infilled with a total mass of 68 million metric tons (~450,000 MT/yr). Rowland and Dietrich (2006) then assumed that all 24 oxbow lakes found on a map are active prior to river alteration, and that all 24 received sediment at a similar rate they established. The loss of suspended sediment to lakes would be 4% of estimated pre-1963 Mississippi load (Rowland and Dietrich, 2006; Keown et al. 1986). The creation of empirical meandering river and oxbow lake relationships is therefore not only important to understanding floodplain architecture, but can also aid sedimentary and marine Geologists who want to better understand, analyze, and predict the global flux of sediment to the ocean.
Chapter 9 Conclusions

The study presented allows for a comparison of geographically disperse rivers. The rivers are in turn comprised of a variety of end-member fluvial systems (i.e. fluvial and tidally dominated). The preceding analysis describes a method to analyze fluvial longitudinal profiles and their floodplain architecture using the high resolution SRTM Digital Elevation Model.

The method to extract the longitudinal profile of a river is tedious to implement due to the necessity to manually trace a river’s centerline. The longitudinal profile depended profoundly on the existence of anthropogenic features (i.e. dams and water diversions), as such limiting of some of the analysis on river equilibrium state. This procedure will allow further research to describe and characterize large-scale fluvial systems throughout the globe. With the fine resolution of SRTM and SWBD it was expected that both large and small river’s would have been resolved. The Kikori and Indus Rivers proved to be a challenge when attempting to extract the water surface elevation for the longitudinal profiles due to their narrow widths. Until resolutions previously available for cost are made free (SRTM 1 arc-sec) or new datasets up-and-coming are captured (Surface Water Ocean Topography), rivers smaller than 180-meters wide are out of scope for this methodology.

Longitudinal profiles of the 16 rivers illustrate that concavity varies greatly across the selected scope of river systems, with the trend fits showing uneven profile lines often
containing local convexities. The process reveals undeniable proof that a river’s longitudinal profile aids in determining the main channel slope and channel planform sinuosity. An empirical relationship of the 16 selected rivers was established between the main active channel’s slope and planform sinuosity, demonstrating that the slope explains the variance of sinuosity. The relationship demonstrates that a high slope yields a low sinuosity while a low slope produces a high sinuosity. While the slope and sinuosity regression analysis shows a strong correlation at the entire river longitudinal profile scale, the correlation decreases when comparing the 5-meander reach scale.

Neotectonics and lithology in turn affect channel slope and sinuosity. The Mississippi and Niger River’s largest deviations in channel sinuosity correlate with growth-fault bound regions within the delta. Strike-slip fault scarps along the active main channel of the Magdalena River cause an abrupt change in river gradient. Bedrock constrictions bound the lateral movement of the rivers, due to the constriction of narrow container valleys. Bedrock transitions are apparent in the longitudinal profile of the Magdalena River, distinguished through an increase in slope.

The inter-floodplain variability of potential bedload flux for the Mississippi and Magdalena Rivers illustrated that numerous sources and sinks exist within the coastal lowlands. The calculations of bedload transport rates require just two major parameters: the slope of the river as derived from SRTM data and observed water discharge. Applying this simple technique predicted bedload fluxes within 85-90% of bedload fluxes found from field studies.
Another important find in this study was the threshold value of 1.6 for the main active channel’s planform sinuosity. Rivers found above the 1.6 threshold sinuosity are all meandering. Meandering rivers are an important end-member in the classification of fluvial systems. Establishing this threshold presents the ability to understand these fluvial systems in their own light.

For example, in Appendix 1, migration rates of the meandering Indus River are established using multiple river centerlines corresponding to different dates. The migration rate for the Indus is found to be higher in the fluvial dominated reach, and less in the tidal dominated reach. The migration rates change over time and are a combination of steady state processes and high intensity flood processes. Another example, Appendix 2, examines the spatial relationship between meandering rivers (the Fly and Mississippi Rivers) and oxbow lakes found throughout their floodplains. Regression analysis reveals a strong relationship for cumulative oxbow lake area with distance from the river’s main channel in each fluvial basin, scaling between the two systems. Another parallel drawn between the Fly and Mississippi River systems is the infill rate and migration rate illustrate a ceiling on the uppermost age of an oxbow lake before it is either infilled or reworked.

The methodology should prove particularly useful when combined with the access of auxiliary data sets (i.e. water discharge, sediment discharge, geology, and fault) in the United States, and with future published and easily attained data for worldwide locations.
The methods would be greatly improved with the access to a global geological dataset. OneGeology is presently undertaking the task to create such a dataset, however, in its present form the coarse resolution would not be applicable to this study. That stated, data similar to the resolution and quality to that used within this study should allow for a similar analysis (comparing the river’s slope control on sinuosity) to the entire encompassing reaches of the SRTM DEM.

With geologic and hydrologic data still lacking for many regions of the world, the passive remote sensing technique methods outlined herein should facilitate a greater understanding of fluvial systems lacking detailed hydrological data of a river’s physical parameters (slope, sinuosity, width, etc.). Expansive global datasets would allow for the expansion of the geographic scope of this study. Encompassing as many rivers and regions as possible could reveal a number of empirical relationships. The creation of truly predictive relationships, through empirical analysis, would require descriptions of any important hydrological parameter suspected to constrain slope and sinuosity of a river.

The techniques outlined herein represent two contributions to the field of fluvial morphology. The first is a refined method in which to examine channel slope and planform sinuosity with remotely sensed data. This benefits the field in the ability to apply field techniques remotely without interfering directly with the system and at cheaper cost. The second contribution is the starting point for empirical relationships to be established for fluvial systems across geographically diverse regions using Geographic
Information Systems. This provides insight to examining techniques for specific classifications of fluvial systems established in this study.
Acknowledgments

References


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Appendix A

This appendix includes some of the fluvial processing routines written in the Python Programming Language. The scripts were written during this study to assist in the processing of the SRTM and SWBD datasets. The routines included here were instrumental in the calculation of lateral migration (and river width) using shapefiles.

These programs should be easily translatable into other object oriented programming languages. Many of the commands below are unique to Python, but other languages have analogous commands. Anything found behind pounds (#) are commented regions. It is also important to note that these routines call in the ArcGIS Geoprocessor, which is essential in most of the functionality of these scripts.

```python
# Create Perpendicular Lines for all line segments within a shapefile
try:
    # Import system modules
    import sys, string, os, arcgisscripting
    gp = arcgisscripting.create()

    # Create the Geoprocessor object
    # Local variables...
    inputlines = r"C:\GIS\work\Yellow\swbd\Yellow_cl_UTMp.shp"

    distance = 1900
    # Where the new perpendicular lines will be written
```

# This is how far out the perpendicular line will be created on either side.
```
textfile = r"C:\GIS\work\Yellow\swbd\output1900.txt"

print "please wait, working on file " + inputlines
print "evaluation distance = " + str(distance)
print "text file location = " + textfile

# Create a text file and write polylines to the first line.
f = open(textfile,'a')
thestring = "polyline\n"
f.writelines(thestring)
f.close()

# Create search cursor
rows = gp.SearchCursor(inputlines)
row = rows.Next()

counter = 0
# start the row iteration
while row:
    # Create the geometry object (where the magic happens)
    feat = row.Shape
    # get coordinate values as lists
    firstpoint = feat.FirstPoint
    lastpoint = feat.LastPoint
    midpoint = feat.Centroid
    # split the lists by the blank space between the coordinate pairs
    firstpoint = firstpoint.split(" ")
    lastpoint = lastpoint.split(" ")
    midpoint = midpoint.split(" ")
    # get the x and y values as array positions 0 and 1, and convert them to floating point
    startx = float(firstpoint[0])
    starty = float(firstpoint[1])
    endx = float(lastpoint[0])
    endy = float(lastpoint[1])
    midx = float(midpoint[0])
    midy = float(midpoint[1])

    # if the line is horiz or vert the slope and negative reciprocal will fail
    if starty==endy or startx==endx:
        if starty == endy:
            """"
\[ y_1 = \text{midy} + \text{distance} \]
\[ y_2 = \text{midy} - \text{distance} \]
\[ x_1 = \text{midx} \]
\[ x_2 = \text{midx} \]

if \( \text{startx} == \text{endx} \):
\[ y_1 = \text{midy} \]
\[ y_2 = \text{midy} \]
\[ x_1 = \text{midx} + \text{distance} \]
\[ x_2 = \text{midx} - \text{distance} \]

else:

# get the slope of the line
\[ m = ((\text{starty} - \text{endy})/(\text{startx} - \text{endx})) \]

# get the negative reciprocal,
\[ \text{negativereciprocal} = -1*((\text{startx} - \text{endx})/(\text{starty} - \text{endy})) \]

if \( m > 0 \):

# increase x values, find y
if \( m >= 1 \):
\[ y_1 = \text{negativereciprocal}*(\text{distance}) + \text{midy} \]
\[ y_2 = \text{negativereciprocal}*(-\text{distance}) + \text{midy} \]
\[ x_1 = \text{midx} + \text{distance} \]
\[ x_2 = \text{midx} - \text{distance} \]

# increase y find x
if \( m < 1 \):
\[ y_1 = \text{midy} + \text{distance} \]
\[ y_2 = \text{midy} - \text{distance} \]
\[ x_1 = (\text{distance}/\text{negativereciprocal}) + \text{midx} \]
\[ x_2 = (-\text{distance}/\text{negativereciprocal}) + \text{midx} \]

if \( m < 0 \):

# add to x find y
if \( m >= -1 \):

# add to y find x
\[ y_1 = \text{midy} + \text{distance} \]
\[ y_2 = \text{midy} - \text{distance} \]
\[ x_1 = (\text{distance}/\text{negativereciprocal}) + \text{midx} \]
\[ x_2 = (-\text{distance}/\text{negativereciprocal}) + \text{midx} \]
if m < -1:
    y1 = negativereciprocal*(distance) + midy
    y2 = negativereciprocal*(-distance) + midy
    x1 = midx + distance
    x2 = midx - distance

f = open(textfile,'a')
    thestring = str(counter) + " 0\n" + "0 "+ str(x1)+" "+str(y1) + "\n" + "1 " + str(x2) + "+" + str(y2) +"\n"
f.writelines(thestring)
f.close()
del x1
del x2
del y1
del y2

    counter = counter + 1

    row = rows.Next()

del row
del rows
f = open(textfile,'a')
    thestring = "END"
f.writelines(thestring)
f.close()

except:
    print "Failed"
    gp.AddMessage(gp.GetMessages(2))
    print gp.GetMessages(2)
del gp

del gp

#Ends the script
stopme = raw_input ("your script has finished, press enter to end")

# Create EndPoints (First and Last) for a Polyline or set of Polylines
# -----------------------------------------------------------------------------
try:
# Import system modules
import sys, string, os, arcgisscripting

# Create the Geoprocessor object
gp = arcgisscripting.create()

# Local variables...
# Input line file
inputlines = r"C:\GIS\work\network\Bartholomew1922IndusCopy.shp"

# Output text file (where it is written)
textfile = r"C:\GIS\work\network\OutTrial10.txt"

print "please wait, working on file " + inputlines
print "text file location = " + textfile

# Create a text file and write polylines to the first line.
f = open(textfile,'a')
thestring = "Point
f.writelines(thestring)
f.close()

# Create search cursor
rows = gp.SearchCursor(inputlines)
row = rows.Next()

counter = 0  # Was at 0

# Start the row iteration
while row:
    # Create the geometry object
    feat = row.Shape
    # Get coordinate values as lists
    firstpoint = feat.FirstPoint
    lastpoint = feat.LastPoint
    midpoint = feat.Centroid
    # Split the lists by the blank space between the coordinate pairs
    firstpoint = firstpoint.split(" ")
    lastpoint = lastpoint.split(" ")
    midpoint = midpoint.split(" ")
    # Get the x and y values as array positions 0 and 1, and convert them to floating point
    startx = float(firstpoint[0])
    starty = float(firstpoint[1])
    endx = float(lastpoint[0])
    endy = float(lastpoint[1])
midx = float(midpoint[0])
midy = float(midpoint[1])

y1 = starty
y2 = endy
x1 = startx
x2 = endx

f = open(textfile,'a')
    thestring = str(counter + 1) + " " + str(x1)+ " " +str(y1) + "\n" + str(counter + 2) + 
    " " + str(x2) + " " + str(y2) +"\n"
    f.writelines(thestring)
    f.close()
    del x1
    del x2
    del y1
    del y2
    counter = counter + 2

    row = rows.Next()
    del row
    del rows
    f = open(textfile,'a')
    thestring = "END"
    f.writelines(thestring)
    f.close()
except:
    print "Failed"
    gp.AddMessage(gp.GetMessages(2))
    print gp.GetMessages(2)
    del gp
    del gp

#Ends the script
stopme = raw_input ("your script has finished, press enter to end")
# Create Points Systematically down a Polyline or set of Polylines

```python
import arcgisscripting, math
gp = arcgisscripting.create()
gp.CheckOutExtension("Spatial")
gp.workspace=r"C:\GIS\work\network\Bartholomew1922IndusCopy.shp"

path=gp.workspace

## Inputs required shp file and DEM ### MTH
theme=path+'\Points.shp'
DEM = "C:\GIS\work\network\arcinfo\fltr_grd"

ex = gp.describe(DEM)
gp.extent = ex.extent
Outpath = path

# Long points are the files created for the point file ### MTH
Longpoints = path+'\Longpoints.shp'
Longpntoutput = path+'\Longpntoutput.txt'

### Need these spacing and elevation ### MTH
Elevation = 'zpoints'
XS_spacing = 15 # This is the spacing between each cross-section. The units are in meters, if you are using UTM

# Here I create a surchcursor and determine how many nodes are in the stream segment
cur = gp.searchcursor(theme)
row = cur.next()
geometry = row.shape

# Here I create a linearray and determine how many nodes are in the stream segment
linearray = geometry.getpart(0)
nodes = linearray.count
del row
TotalLength = geometry.length

# Starting Conditions
NumNodes = int((2*XSlen)/XSpacing)
dist = 0
remainder = 0
CumLength = 0
```
count=0
XSxCenterPnts=[] # I am making a list of the x points where I create a cross-section
XSyCenterPnts=[] # I am making a list of the y points where I create a cross-section
TestXnodes=[]
TestYnodes=[]
TestTheta=[]
TestHypotenuse=[]
Testb=[]
Testc=[]
Timesinifloop=[]
Testremainder=[]
Testdistcurrentseg=[]
Testdist=[]
print "Going into for loop"
for i in range(0,node-1):

    XSxPnts=[]
    XSyPnts=[]
    # First I'm going to get the x,y coordinates of the first two nodes in the stream polyline
    # starting at the top of the stream and working downstream
    point1=linearray.getobject(i)
    point2=linearray.getobject(i+1)
    x1=point1.x
    TestXnodes.append(x1)
    y1=point1.y
    TestYnodes.append(y1)
    x2=point2.x
    y2=point2.y
    # Now I calculate the distance between the first two nodes
    # cross-section if not I will just keep track of the distance until I cross the threshold.
    distCurrentSeg=math.sqrt(pow((x2-x1),2)+pow((y2-y1),2))
    Testdistcurrentseg.append(distCurrentSeg)
    CumLength=CumLength+distCurrentSeg

    if dist>XSspacing and CumLength<=TotalLength:
        count +=1
        print count
        Timesinifloop.append(i)

    # Now determine the location where you need to place the cross-section line
    if dist>XSspacing and CumLength<=TotalLength:
        count +=1
        print count
        Timesinifloop.append(i)
pntoutput = Outpath+'\'+XSdata+str(count)+'.txt'
print pntoutput
zpoints = Outpath+'\'+Elevation+str(count)+'.shp'
print zpoints
XSec=gp.createfeatureclass(Outpath, CrossSect+str(count)+".shp", "Point")
print XSec
gp.addfield(XSec, "Distance", "double")
print "Made Cross-section feature class, need to create cross-section data"
#change then the second statement will show the correct x & y coordinates.
print "Getting Geometry"
if math.fabs(x2-x1)==0:#Here I'm asking: Do the x coordinates change, or is it a
vertical line?
    XCenterPnt=x1
    XSxCenterPnts.append(XCenterPnt)
    YCenterPnt=math.fabs(y1-XSspacing)
    XSyCenterPnts.append(YCenterPnt)
else:
    if x2>x1:#I determine whether the second node has larger x and y values here.
        This influences the sign of Cos*XSspacing and Sin*XSspacing arguments
        b=2
    else:
        b=1
    if y2>y1:
        c=2
    else:
        c=1
    Testb.append(b)
    Testc.append(c)
    theta=math.atan(math.fabs(y1-y2)/math.fabs(x2-x1))
    TestTheta.append(theta)
    remainder=dist-XSspacing
    Testremainder.append(remainder)
    hypotenuse=distCurrentSeg-remainder
    TestHypotenuse.append(hypotenuse)
    XCenterPnt=(pow(-1,b))*math.cos(theta)*hypotenuse+x1
    XSxCenterPnts.append(XCenterPnt)
    YCenterPnt=(pow(-1,c))*math.sin(theta)*hypotenuse+y1
    XSyCenterPnts.append(YCenterPnt)
print "The cross-section x,y points have been located on the stream and added to
the list"
    del b,c, theta
cursor = gp.insertcursor(CrossSectCent)
row = cursor.newrow()#create a new row
row.Distance=0
Centpointc=CenterPntArray.getobject(0)
row.shape=Centpointc
row.ID=0
cursor.insertrow(row)
cursor.newrow()
CumDistancebetweenPoints=0
for n in range(1,len(XSxCenterPnts)):
    Centpointc=CenterPntArray.getobject(n-1)
    Centpointd=CenterPntArray.getobject(n)
    row.shape=Centpointd
    row.ID=n
CumDistancebetweenPoints=CumDistancebetweenPoints+math.sqrt(pow((Centpointd.x-Centpointc.x),2)+pow((Centpointd.y-Centpointc.y),2))
row.Distance=CumDistancebetweenPoints
cursor.insertrow(row)
cursor.newrow()
del cursor, row, CumDistancebetweenPoints, XSyCenterPnts, xe1, xe2, ye1, ye2,
XSxPnts, XSyPnts, YCenterPnt, XCenterPnt
gp.ExtractValuesToPoints_sa(CrossSectCent,DEM, Longpoints, "NONE", "ALL")
Longcursor = gp.searchcursor(Longpoints)
Longrow = Longcursor.next()
Longdistlist = []
Longelevlist = []
print "Dumping the Cross-Section values into a text file."
#Populate distance field of zpoints
for t in range(0,len(XSxCenterPnts)-1):
    print "calculating distance for point "+str(t)
    Longrow = Longcursor.next()
    Longelev = Longrow.rastervalu
    Longelevlist.append(Longelev)
    Longdist=Longrow.Distance
    Longdistlist.append(Longdist)
#Creating list of tuples to output to text file
fileloop = range(len(Longelevlist))
Longpntlist = []
print "fileloop is: 
print fileloop
for v in fileloop:
    Longpntlist.append(str(Longdistlist[v])+", ")
    Longpntlist.append(str(Longelevlist[v])+"n")
print "here is pointlist:"
print Longpntlist
del Longcursor, Longrow, XSxCenterPnts, linearray
#Writing output textfile
file = open(Longpntoutput,'w')
file.writelines("distance m, elevation m\n")
file.writelines(Longpntlist)
file.close()
print "output file written!"