The Pennsylvania State University
The Graduate School
College of Earth and Mineral Sciences

THE GROWTH AND EVOLUTION OF RIVER-DOMINATED DELTAS
AND THEIR DISTRIBUTARY NETWORKS

A Dissertation in
Geosciences
by
Douglas A. Edmonds

© 2009 Douglas A. Edmonds

Submitted in Partial Fulfillment
of the Requirements
for the Degree of

Doctor of Philosophy

May 2009
The dissertation of Douglas A. Edmonds was reviewed and approved* by the following:

Rudy Slingerland  
Professor of Geology  
Dissertation Advisor  
Chair of Committee

Eric Kirby  
Associate Professor of Geosciences

David Hill  
Associate Professor of Civil and Environmental Engineering

Richard Alley  
Evan Pugh Professor of Geosciences

Katherine Freeman  
Professor of Geosciences  
Associate Department Head of Graduate Programs and Research

*Signatures are on file in the Graduate School
ABSTRACT

River-dominated deltas are dynamic environments and because they are home to a significant fraction of the world’s population we need to understand their growth and evolution so their behavior can be predicted and hazards can be mitigated. Here, using a combination of numerical modeling, physical experiments, and field data, I investigate the processes that participate in the growth and evolution of river-dominated delta channel networks. Using physical experiments I document that deltaic avulsions are caused by an upstream migrating wave of sedimentation that is triggered by a stagnated river mouth bar. An avulsion occurs at the levee location where the greatest average shear stress has been exerted for the longest time. The subsequent evolution of the delta network is a function of the configuration and stability of the individual bifurcations that divide water and sediment. Numerical modeling experiments show that the discharge ratio in the downstream bifurcate channels is a function of the Shields number in the upstream channel. There are two equilibrium functions where one defines symmetrical configurations (equal partitioning of discharge), while the other two define asymmetrical configurations (unequal partitioning of discharge). A network of equilibrium bifurcations is stable to perturbations. Using numerical experiments and field data I show that delta networks are generally stable to perturbations in the form of a closure of a bifurcate channel. Interestingly though, the effect of that perturbation redistributes the water and sediment fluxes throughout the delta, which has the potential to change the long term evolution of a delta. Most notably, these results have implications for engineering fluvial systems. For example, in the 1960s it was thought that all the water from the Mississippi River would flow down the Atchafalaya River. To stabilize this bifurcation a control structure was built to regulate the discharge distribution. The results in this thesis suggest that perhaps a stable bifurcation could have been designed without the aide of control structures.
TABLE OF CONTENTS

LIST OF FIGURES..................................................................................................................... vii

LIST OF TABLES......................................................................................................................... x

ACKNOWLEDGEMENTS.............................................................................................................. xi

Chapter 1: Introduction........................................................................................................... 1

1.1 Historical significance of deltas ....................................................................................... 2
1.2 Early sedimentological studies of deltas ......................................................................... 3
1.3 Constraining delta morphology ....................................................................................... 4
1.4 Early numerical modeling of river-dominated deltas ................................................... 7
1.5 Morphodynamic modeling and the goals of this research ............................................. 8
1.6 Summary of dissertation chapters .................................................................................. 9

Chapter 2: Predicting delta avulsions: Implications for coastal wetland restoration. 13

Abstract ................................................................................................................................ 14

2.1 Introduction ..................................................................................................................... 15

2.2 Hypothesis and methodology ......................................................................................... 16

2.3 The avulsion cycle in deltas........................................................................................... 20

2.3.1 River mouth bar growth and stagnation ................................................................. 20

2.3.3 Morphodynamic backwater causes increased overbank flow ................................... 23

2.3.4 Predicting avulsion location .................................................................................... 25

2.4 Application to real deltas ............................................................................................... 28

2.4.1 Prediction of avulsion location in real deltas .......................................................... 29

2.4.2 Prediction of wetland growth in real deltas ............................................................. 30

2.4.3 How does the avulsion process interact with other network forming processes? ...... 31

2.5 Conclusions .................................................................................................................... 31

Chapter 3: Stability of delta distributary networks and their bifurcations ........................ 33
LIST OF FIGURES

Figure 1-1 MODIS Satellite image of the Nile delta in 2000.................................................. 2
Figure 1-2 Tri-partite classification of delta morphology......................................................... 6

Figure 2-1 Image of the Mississippi delta and overhead photo of an experimental delta
 created with the cohesive sediment mixture........................................................................... 17

Figure 2-2 Normal probability plots of channel lengths nondimensionalized by their
 widths; and channel top widths nondimensionalized by the channel width at the
 delta head for experimental and real deltas........................................................................ 18

Figure 2-3 Precision of the stereo camera............................................................................... 19

Figure 2-4 Overhead photographs showing the avulsion cycle for experiment DL2...... 21

Figure 2-5 The morphodynamic backwater occurs in all experiments in this study.. ... 22

Figure 2-6 The morphodynamic backwater increases overbank flow as it moves
 upstream......................................................................................................................... 24

Figure 2-7 The avulsion occurs at the location of $I_{\text{max}}$.................................................. 27

Figure 2-8 The avulsion location through time in DL5......................................................... 28

Figure 3-1 Two examples of distributary deltas with bifurcating channel networks. ...... 35

Figure 3-2 Numerical grid ................................................................................................... 44

Figure 3-3 Two different initial bed ramp configurations tested in this study ............... 46

Figure 3-4 Evolution of the discharge ratio for different initial $\Theta_r$................................. 48

Figure 3-5 Examples of symmetrical and asymmetrical water surface elevation and bed
topography at equilibrium as computed by Delft3D......................................................... 51

Figure 3-6 Equilibrium water surface profiles for an asymmetric bifurcation................. 54

Figure 3-7 Equilibrium diagram for fine-grained, cohesive deltaic bifurcations......... 55

Figure 3-8 For a given $\Theta_r$, an increase in bed ramp height ($\eta$) increases the water
surface at the entrance to channel $b$.................................................................................. 56
Figure 3-9 In equilibrium bifurcations computed in this study there is an inverse relationship between bedload transport and discharge in channel b ................................. 56

Figure 3-10 Evolution of a perturbed equilibrium bed topography .............................................. 59

Figure 3-11 River bed topography of natural bifurcations on the Mossy delta, Saskatchewan, Canada ........................................................................................................ 62

Figure 3-12 Water surface and bed elevations on a bifurcation in the Mossy delta ............. 63

Figure 3-13 Stable bifurcations from the Mossy delta generally plot in the stable, equilibrium space predicted by Delft3D ...................................................................... 64

Figure 4-1  Numerical grid overlain on 2003 composite aerial photograph of the Mossy Delta, Saskatchewan, Canada ............................................................................... 76

Figure 4-2 Calculated versus observed discharge in various reaches of the Mossy delta.. ................................................................................................................................. 80

Figure 4-3 Mossy delta reaches equilibrium after 17 years of computation ................. 88

Figure 4-4 Water surface elevations calculated by the model compare well with field data..................................................................................................................................... 90

Figure 4-5 Predicted equilibrium water depths in distributaries of the Mossy delta .... 91

Figure 4-6 All nine perturbation experiments come to a new equilibrium after a channel is closed..................................................................................................................... 94

Figure 4-7 Closing one channel causes reorganization of the water discharge throughout the delta network ........................................................................................................ 97

Figure 4-8 Water surface elevation adjustment after a channel is closed ....................... 98

Figure 4-9 After a closure, fractional change in discharge ratio in affected bifurcations positively correlates with the fractional change in Shields stress ....................... 101

Figure 4-10 The percentage of sediment flux entering the head of the delta that is discharged out of the north and south halves at equilibrium ................................. 102

Figure 4-11 Aerial image of the Lena River delta from Landsat 7 satellite ..................... 103
Figure A-1  Location map for field data. ................................................................. 125
Figure A-2  Data collected in 2006 for bifurcation 1.............................................. 126
Figure A-3  Data collected in 2007 for bifurcation 1.............................................. 127
Figure A-4  Data collected in 2006 for bifurcation 2.............................................. 128
Figure A-5  Data collected in 2006 for bifurcation 3.............................................. 129
Figure A-6  Data collected in 2007 for bifurcation 3.............................................. 130
Figure A-7  Data collected in 2006 for bifurcation 4.............................................. 131
Figure A-8  Data collected in 2006 for bifurcation 5.............................................. 132
Figure A-9  Data collected in 2007 for bifurcation 6.............................................. 133
Figure A-10 Data collected in 2007 for bifurcation 7............................................ 134
Figure A-11 Data collected in 2006 for bifurcation 9............................................ 135
Figure A-12 Composite water surface elevation map from the Mossy delta, Saskatchewan, Canada................................................................. 136

Figure B-1 Estimated error in $I$ for DL2...................................................... 138
Figure B-2 Estimated error in $I$ for DL4...................................................... 139
Figure B-3 Estimated error in $I$ for DL5...................................................... 140
Figure B-4 Estimated error in $I$ for DL9...................................................... 141
Figure B-5 Estimated error in $I$ for DL12..................................................... 142
LIST OF TABLES

Table 2-1 List of the five physical experiments....................................................... 25

Table 4-1 List of the nine perturbation experiments................................................... 93
ACKNOWLEDGEMENTS

It is difficult to say how I became interested in geology, or river deltas for that matter. But it is probably accurate to say that my interest was the result of a childhood spent mucking through rivers and creeks, and by the time I enrolled at Penn State for graduate school I only knew that I wanted to study rivers in some capacity. With a photograph of a strikingly symmetric river delta and some statistics, Rudy Slingerland focused my interests onto the problem of delta morphology. As my advisor, Rudy has been an unending source of support and advice and most importantly friendship. Without his friendship I might never have enjoyed myself as much or laughed as hard. The same can also be said for my other committee members, Eric Kirby, David Hill, and Richard Alley who selflessly gave of their time and provided thoughtful guidance.

I am also indebted to David Hoyal, Ben Sheets, and Roger Bloch who have played important roles in my maturation as a scientist. Furthermore, I am grateful to John Bridge, Jim Best, David Janesko, Frank Klein, Katy McGuire, Dan Parsons, Norm Smith, and Arjan Reesnik who were part of the Cumberland Marshes field team during the summers of 2006 and 2007.

An immeasurable benefit of Penn State is being a part of the larger graduate student community. The list of people from that community to whom I am indebted is too long to enumerate. In particular though, there are some who have been instrumental in helping me think about my career and my research: Charlie Angerman, Jon Barton, James Bonelli, Brooke Fambrough, Evan Goldstein, Dave Greene, Nathan Harkins, Brian LeVay, Scott Miller, Dan Peterson, Andy Rathbun, Tyrone Rooney, Rob Selover, and Dave Vacco. I would like to personally thank Scott Miller for always being an available sounding board and guiding me through my first few years of graduate school.

I am also thankful to my parents and family who have supported me the entire way. Of course, I owe my most sincere thank you to my loving fiancé who understood why I was gone in the early mornings and not around in the evenings. Thank you for always supporting me; it is to you that I owe everything.
Mark Twain on the usefulness of scientific data:

Now, if I wanted to be one of those ponderous scientific people, and 'let on' to prove what had occurred in the remote past by what had occurred in a given time in the recent past, or what will occur in the far future by what has occurred in late years, what an opportunity is here! Geology never had such a chance, nor such exact data to argue from! ... Please observe:--

In the space of one hundred and seventy-six years the Lower Mississippi River has shortened itself two hundred and forty-two miles [due to meander cutoffs]. That is an average of a trifle over one mile and a third per year. Therefore, any calm person, who is not blind or idiotic, can see that in the Old Oolitic Silurian Period, just a million years ago next November, the Lower Mississippi River was upwards of one million three hundred thousand miles long, and stuck out over the Gulf of Mexico like a fishing-rod. And by the same token any person can see that seven hundred and forty-two years from now the Lower Mississippi will be only a mile and three-quarters long, and Cairo and New Orleans will have joined their streets together, and be plodding comfortably along under a single mayor and a mutual board of aldermen. There is something fascinating about science. One gets such wholesale returns of conjecture out of such a trifling investment of fact.

--p. 106, Life on the Mississippi, 1883
Chapter 1: Introduction
1.1 Historical significance of deltas

Deltas have long been a fascination of humans. This fascination arose from the fundamental importance of deltaic plains in early civilizations: their fertile land was an agricultural asset and their proximity to the coastline provided ample opportunities for trading, exploring, and utilizing maritime resources. Given this, deltas captured the attention of many ancient scholars, such as Homer, Plato, and Aristotle. But it was the Greek historian Herodotus, who in approximately 450 BC first used the term “delta” to describe the shape of the sedimentary deposit at the mouth of the Nile River because it resembled the Greek letter Δ (Figure 1-1). Much later, Pliny the Elder, in approximately 70 AD, recognized that the Nile delta was composed of a network of seven river channel branches that created the triangular form. Considering the importance of delta systems to ancient civilizations, it is not surprising that these scholars contributed, perhaps, the first recorded descriptions of delta morphology and ideas regarding the origin of that morphology.

Figure 1-1 MODIS Satellite image of the Nile delta in 2000. The deposits of the Nile river resemble an upside down “Δ”. The image is from NASA.
1.2 Early sedimentological studies of deltas

Since then, our knowledge about delta morphology has largely been supported by the efforts of geologists studying delta deposits in rock outcrops. The first published sedimentological description of a delta came from the great G.K. Gilbert [Gilbert, 1885]. Gilbert studied the deltas that formed in glacial Lake Bonneville, Utah and recognized that they generally contained three units: a top set, fore set, and bottom set. These three terms, which describe the environments of deposition within a delta, are still in use today. Barrell [1912] extended Gilbert’s ideas to the Catskill delta, a much larger system, and proposed criteria for the recognition of deltaic deposits in the ancient rock record. The seminal work of these and other early 20th century geologists led to the insight that delta deposits were superb oil reservoirs due to the thick accumulation of organic matter and clastic sediment that occurs in the deltaic environment.

With the technological and intellectual breakthroughs of the early 20th century, geologists were able to more successfully describe the sedimentary architecture of ancient deltas [Bhattacharya, 2006]. These better descriptions of ancient delta systems sparked the important realization that not all deltas are created equal; there is a range of delta morphologies each with distinct sedimentological assemblages. An important next step in understanding delta systems was to generate a causal link between their morphology and sedimentology. This also had practical value because in sedimentary basins with limited information, if the morphology could be constrained, then the sedimentological assemblage could be predicted with little additional exploration and cost.
1.3 Constraining delta morphology

It was difficult to constrain the controls on delta morphology for early 20th century geologists because their primary dataset was the geologic record. The geologic record did not always provide a complete picture of delta morphology because the exposure of the record is limited, and not all events are preserved in the geologic record. To overcome this limitation geologists shifted their focus and began studying the geomorphology of modern deltas. The geomorphologic approach was aided by the availability of aerial and satellite imagery, which, for the first time, made a comprehensive study of modern delta morphology possible. From analysis of aerial photography, maps, and field studies, delta morphology was hypothesized to be a result of the interaction among the forces of rivers, waves, and tides (Figure 1-2) [Galloway, 1975]. Researchers recognized that deltas dominated by fluvial forces were composed of intricate channel networks with many river branches (see river-dominated deltas in Figure 1-2). On the other hand, if the forces of waves dominated, the channel network was suppressed, resulting in a prograding channel and beach-face complex (see wave-dominated deltas in Figure 1-2) [Bhattacharya, 2006]. The dominance of tides tended to complicate things and produced a dendritic channel network (see tide-dominated deltas in Figure 1-2). Galloway’s tri-partite classification of deltas may be too simplistic [e.g., Wright, 1977; Orton and Reading, 1993; Heap, et al., 2004; Bhattacharya, 2006], but it served as a significant step forward in understanding delta morphology and a remarkable demonstration of the link between the dominance of certain processes and deltaic morphology.

Such a strong link between process and morphology exists in delta systems
because there is a unique mapping between process and the resultant form. For example, a simple survey of the world’s river-dominated deltas reveals that most have branching, or bifurcating, channels and further inspection shows that the majority of their bifurcations distribute flow and sediment asymmetrically. Researchers recognized that the fundamental process controlling these patterns was channel bifurcation at the river mouth [Bates, 1953; Wright, 1977] and therefore understanding the morphology of river-dominated deltas reduced to understanding the processes at the river mouth. This was an important outcome because it demonstrated that there is value in understanding the process-morphology connection in deltaic systems.
Figure 1-2 Tri-partite classification of delta morphology [after Galloway, 1975]. Deltas dominated by rivers, waves, or tides have different morphologies in terms of the number of active channels, branching pattern, and shoreline morphology. Images taken by LANDSAT 7 satellite and are from the NASA website.
1.4 Early numerical modeling of river-dominated deltas

The advancement of numerical modeling provided a useful tool for elucidating the relationship between process and morphology in deltas because geologists simulated delta processes and “watched” the morphology of deltas develop. The modeling of river-dominated deltas was a reasonable first challenge to geomorphologists because the equations for flow and sediment transport in the absence of waves and tides were fairly well constrained. The earliest numerical models of river-dominated deltas were based on the simplified physics of water flow and sediment transport and simulated a sediment-laden river entering a standing body of water. From there scientists tested their curiosity by changing boundary conditions and observing the patterns of sediment deposition seaward of the river mouth [Bonham-Carter and Sutherland, 1968; Waldrop and Farmer, 1973; Wang, 1984]. Given certain hydrodynamic boundary conditions a triangular shaped sediment bar, also called a river mouth or distributary mouth bar, is deposited a characteristic distance seaward of the river mouth. Many researchers hypothesized that these river mouth bars created incipient bifurcations. However, it was not clear if this was true because the models did not predict how a bar evolved into a bifurcation.

Early models could not predict how a bar evolved into a bifurcation because the models were conceptualized with a “one-way” connection between process and morphology; the information from the process created the morphology, but the information from the newly created morphology did not in turn, influence the process. While these “one-way” models were illustrative they could not reach the ultimate goal of understanding how the process produced the morphology.
1.5 Morphodynamic modeling and the goals of this research

As computing power increased and numerical models of flow and sediment transport became more sophisticated, a new modeling strategy emerged with a “two-way” connection between process and morphology. Morphodynamic modeling, as it is now known, holds an advantage over the previous numerical models because the algorithms for flow and sediment transport are fully coupled; the changes in bed topography cause changes in the flow field in real time allowing for prediction of the complete dynamical evolution of a system. With this technique it is now possible to approach some of the questions that earlier researchers could not answer. At present the application of morphodynamic modeling to deltas has been minimal, and therefore many questions remain regarding the growth and evolution of river-dominated delta networks. Three basic questions arise: 1) What processes participate in the formation of river-dominated delta channel networks; 2) Once formed, what are the equilibrium states of the delta channel network; and 3) How stable are those states to perturbations?

Answering these questions is important because if the processes responsible for river-dominated delta formation are better understood then our ability to predict the behavior of deltaic environments will be more successful. Predicting delta behavior is important because, after all, human fascination with deltas has not ceased—currently 25% of the world’s population lives in and around deltaic plains [Syvitski, et al., 2005b].—and this is especially troubling in the context of climate change and sea level rise, which will certainly have a substantial effect on deltaic environments [McCarthy, et al., 2001]. Therefore a better understanding of the processes responsible for delta
growth and evolution and an understanding of how deltas respond to perturbations will take an important first step toward mitigating future hazards.

1.6 Summary of dissertation chapters

This dissertation consists of three chapters (2-4) that are written as stand-alone papers for journal publication. These chapters are linked by my motivation to better understand the processes and feedbacks that create and maintain river-dominated delta channel networks.

Chapter 2 investigates the processes that form river-dominated delta networks. Previous workers [Edmonds and Slingerland, 2007 and references therein] have demonstrated that a river mouth bar forms offshore when a river meets a standing body of water. The river mouth bar becomes the location of channel bifurcation and a delta network forms as this process repeats in time and space. However, this is not the whole story because recent studies [Coleman, 1988; Swenson, 2005; Edmonds and Slingerland, 2007; Jerolmack and Swenson, 2007; Hoyal and Sheets, in press] have shown that the process of channel avulsion also contributes to the formation of channel networks. Currently little is known about the extent to which avulsions contribute to delta network construction and the mechanics of how those avulsions occur. The goals of this chapter are to document and quantify how the process of avulsion contributes to delta construction. Using physical experiments, I investigate the controls on the timing and location of avulsions in delta networks. I conducted physical experiments of eight
different delta lobes and collected data on the evolving bed and water surface
topography each. From this dataset I demonstrate that avulsions in deltas are triggered
by a river mouth bar growing at the shoreline. The mouth bar triggers a wave of bed
aggradation moving upstream that increase cross-levee flows and bed shear stresses.
An avulsion occurs as a time-dependent failure of the levee where the largest average
bed shear stress has been applied for the longest time. This work establishes one
mechanism for how avulsions in deltas occur and can contribute to the formation of the
delta network. This chapter is in press at the journal Geology. D. Edmonds conceived
the project goals, conducted the experiments at Exxon Mobil Upstream Research
Company with the assistance of David Hoyal and Ben Sheet, conducted data analysis,
and wrote the paper. D. Hoyal, B. Sheets, and R. Slingerland critically evaluated the
results and provided editorial comments.

After a delta network forms, its equilibrium state depends upon the equilibrium
state of the bifurcations because they distribute water and sediment throughout the
delta. Therefore, Chapter 3 focuses on the equilibrium conditions of bifurcations and
the processes and feedbacks that keep them stable. A survey of bifurcations within the
world’s deltas reveals a surprising fact: bifurcations on average distribute water and
sediment asymmetrically. This chapter explores what processes create asymmetrical
bifurcations and the feedbacks that keep them stable. I conducted numerical modeling
with Delft3D and found that there are three equilibrium configurations for asymmetrical
delta bifurcations: symmetrical (equal partitioning of discharge), while the other two
define asymmetrical configurations (unequal partitioning of discharge). I then
demonstrated that when the symmetrical equilibrium configuration is perturbed it finds
a new asymmetrical solution, whereas when the asymmetrical equilibrium configuration is perturbed it returns to its original state. This suggested that the asymmetrical bifurcations are prevalent in natural systems because of myriad natural perturbations that eventually force the symmetrical bifurcation asymmetrical. Finally, I validated the numerical model results using field data from the Mossy delta, Saskatchewan, CA. This chapter has been published in the journal Water Resources Research [Edmonds and Slingerland, 2008]. D. Edmonds conceived the project goals and design, conducted the numerical modeling experiments, analyzed the data, and wrote the paper. R. Slingerland assisted in field data collection and critical analysis of the results, and provided editorial comments.

Chapter 4 takes a broader approach and examines how an entire network of bifurcations responds to perturbations. While Chapter 3 showed that individual bifurcations are stable to perturbations, it is not clear how an entire network of bifurcations will respond to a larger perturbation, such as the closing of a channel. To answer this question, I conducted numerical modeling experiments with Delft3D. The experiments were designed to replicate the Mossy delta in Saskatchewan, CA. The first set of numerical experiments was designed to determine if the Mossy delta is in equilibrium. Field data collected over two seasons were used as initial and boundary conditions. The model then computed forward in time until it reached equilibrium, which occurs when there are no changes in the bed topography in the model during one time step. Results show that the Mossy delta is in equilibrium with the incoming water and sediment discharge. At equilibrium the calculated discharge distribution through the network and the water surface topography compares favorably with data collected in
The second set of experiments started from the equilibrium condition of the Mossy delta and then perturbed that condition. I conducted nine experiments where I forced the closure of a bifurcate channel in each experiment. The channels ranged in bifurcation order and discharge. Results show that when a channel is closed the effects are felt throughout the channel network. The extent of that effect is a function of the discharge of the closed channel and the proximity of other bifurcations and in some cases the effect may alter the long term evolution of the delta. D. Edmonds conceived the project goals and design, conducted the numerical modeling experiments, analyzed the data, and wrote the paper. R. Slingerland assisted in field data collection and critical analysis of the results, and provided editorial comments. J. Bridge, J. Best, D. Parsons, and N. Smith assisted in data collection.

Together these studies demonstrate how river-dominated deltas grow and evolve through time. They extend the body of knowledge on river-dominated deltas by describing and quantifying how certain processes contribute to the growth and evolution of these systems. The following chapters will show that these processes should be considered when trying to understand and predict the morphology of the river-dominated deltas.
Chapter 2: Predicting delta avulsions: Implications for coastal wetland restoration

Doug Edmonds\textsuperscript{1}\textsuperscript{*}
David Hoyal\textsuperscript{2}
Ben Sheets\textsuperscript{3}
Rudy Slingerland\textsuperscript{1}

\textsuperscript{1}The Pennsylvania State University, Department of Geosciences, University Park, PA
\textsuperscript{2}Exxon Mobil Upstream Research Company, Houston, TX,
\textsuperscript{3}University of Washington, School of Oceanography, Seattle, WA

\textsuperscript{*}Corresponding author

*Corresponding author

submitted to GEOLOGY
Received 17 December 2008
Accepted 10 March 2009
Abstract

River deltas create new wetlands through a continuous cycle of delta lobe extension, avulsion, and abandonment, but the mechanics and timing of this cycle are poorly understood. Here we use physical experiments to quantitatively define one type of cycle for river-dominated deltas. The cycle begins as a distributary channel and its river mouth bar prograde basinward. Eventually the mouth bar reaches a critical size and stops prograding. The stagnated mouth bar triggers a wave of bed aggradation that moves upstream and increases overbank flows and bed shear stress on the levees. An avulsion occurs as a time-dependent failure of the levee where the largest average bed shear stress has been applied for the longest time ($R^2 = 0.93$). These results provide a guide for predicting the growth of intra-delta lobes, which can be used to engineer the creation of new wetlands within the delta channel network and improve stratigraphic models of deltas.
2.1 Introduction

Given the importance of wetlands in protecting coastlines from storm surges [Danielsen, et al., 2005; Costanza, et al., 2006; Day Jr, et al., 2007; Barbier, et al., 2008] and maintaining a healthy ecosystem, there is considerable interest in coastal wetland restoration in the world’s deltas [Michener, et al., 1997; Smit, et al., 1997; Valdemoro, et al., 2007]. Restoration plans [U.S Army Corps of Engineers, 2004; Reed and Wilson, 2004; Costanza, et al., 2006] commonly advocate a philosophy of restoring and taking advantage of the natural processes that create wetlands. Most coastal wetlands are naturally created within the active delta channel network [Coleman, 1988; Day, et al., 2000] as channels at the shoreline prograde basinward, bifurcate around river mouth bars (RMBs) [Bates, 1953; Wright, 1977; van Heerden and Roberts, 1988; Edmonds and Slingerland, 2007], and avulse to new locations [Coleman, 1988; Swenson, 2005; Edmonds and Slingerland, 2007; Jerolmack and Swenson, 2007; Hoyal and Sheets, in press]. The formation of bifurcations can already be predicted [Edmonds and Slingerland, 2007], but to restore and take advantage of the complete cycle in wetland restoration we need to understand what factors control the timing and location of deltaic avulsions.

Delta avulsions occur across a variety of time and space scales. For example, on the Mississippi, delta lobe switching originates at the apex of the delta approximately every thousand years [Coleman, et al., 1998], whereas intra-delta lobe switching occurs within the active channel network approximately every hundred years [Coleman and Gagliano, 1964; Coleman, 1988]. Hoyal and Sheets [in press] suggested that the latter class of delta avulsions is controlled by downstream processes rather than upstream
processes. In experimental deltas, they observed that an upstream migrating flow disturbance creates flooding, which leads to avulsion. However, their measurement technique (dye and overhead photos) did not allow quantification of how the flow disturbance propagates upstream and causes an avulsion, or when and where the avulsion occurs. Here, we use novel experimental techniques to characterize the evolving bed and water surface in experimental deltas. We present a clear description of avulsion mechanics in intra-delta lobes and demonstrate for the first time that the location and timing of downstream-controlled avulsions are predictable.

2.2 Hypothesis and methodology

We test the hypothesis that intra-delta lobe avulsions in homopycnal, river-dominated deltas are the result of two processes—distributary channel lengthening (the setup) and the growth of RMBs (the trigger). As the distributary channel within the intra-delta lobe lengthens, a RMB forms at its mouth and is recycled basinward. Eventually, the RMB stagnates and triggers a period of increased bed aggradation and overbank flow, which, in turn, leads to avulsion.
To test this hypothesis, we conducted physical scale modeling experiments of delta systems in a 3 by 5 meter tank of standing water with no allogenic forcing. The boundary conditions consisted of steady, uniform sediment feed rate (18.2 g min⁻¹) and water discharge (10 L min⁻¹) entering into a basin (~4 cm depth) through a constant width slot (0.038 m). The sediment mixture ranges from bentonite clay to coarse sand, and is combined with stabilizing polymer to reproduce the dynamics of fine-grained, cohesive deltas. The processes in the experimental deltas are similar to those in real deltas because in planview they look similar (Figure 2-1), and the distributions of channel lengths and widths that compose the delta network are similar in each case (Figure 2-2). The deltas created in this study are constantly at or above bankfull discharge and therefore represent evolution over scores of floods.

Figure 2-1 (A) 2001 Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) image of the Mississippi delta courtesy of USGS National Center for EROS and NASA Landsat Project Science Office. (B) Overhead photo of an experimental delta created with the cohesive sediment mixture. White spots are foam on the water surface.
Figure 2-2  Normal probability plots of (A) channel lengths nondimensionalized by their widths; and (B) channel top widths nondimensionalized by the channel width at the delta head.  (C) Overhead photo of an experimental delta showing the variable definitions. The channel lengths and widths for the experimental and real deltas have similar distributions, means, and standard deviations. This suggests that the channel network-forming processes in experimental deltas are similar to the processes in real deltas. The data for the real deltas are from 11 river dominated deltas throughout the world (Edmonds and Slingerland, 2007). The data for the experimental deltas are from Agg2 experimental delta in (Hoyal and Sheets, 2008).
We produced four deltas under identical boundary conditions and collected data on ten intra-delta lobes (labeled DL1, 2, 4, 5, 7-12). On each intra-delta lobe we used a StarCam, a commercially available stereo camera with millimeter-scale horizontal and vertical resolution (Figure 2-3); to record the bed and water surface topographies at twenty minute intervals until an avulsion occurred. To collect bed topography we turned off the water and sediment mixture entering the basin and then scanned the bed surface. After scanning we turned the water on, allowed the system to re-equilibrate, and injected titanium dioxide to make the water opaque. We scanned the surface again, this time recording the water surface topography both within the channel and overbank. Of the ten lobes two (DL10, 11) will not be considered here because during the experiment the channel became entrenched against the tank wall and did not avulse. For a more detailed discussion of the methodology and scaling issues see Hoyal and Sheets [in press].

Figure 2-3 (a) Seven replicate scans of bed and water surface in the image were taken to test the precision of the StarCam. (b) The replicate scans show good agreement among one another for an arbitrary selected channel cross-section. The bed surface scans have a standard deviation of approximately 0.5 millimeter for the selected cross-section while the water surface scans have a standard deviation of approximately 1 millimeter. The water surface scans have a higher standard deviation because they are taken while the water is flowing.
2.3 The avulsion cycle in deltas

Analysis of time-series photography and topography of the eight lobes shows a common sequence of morphodynamic events leading to avulsion and lobe abandonment (Figure 2-4).

2.3.1 River mouth bar growth and stagnation

Initially, the distributary channel and the RMB prograde with little to no bed aggradation along the channel (Figure 2-5a, $t/t_a = 0$ to 0.6). During progradation the RMB enlarges, which eventually leads to stagnation, aggradation of the bar to sea level, and splitting of the flow. The distance of RMB progradation, and therefore the length of the newly created intra-delta lobe, is proportional to $M$, the jet momentum flux at the channel mouth, and inversely proportional to grain size to approximately the one-fifth power [Edmons and Slingerland, 2007].
Figure 2-4 Overhead photographs showing the avulsion cycle for experiment DL2. The water is dyed pink to show where flow is channelized and overbank. (a) Initially, the distributary channel forms and then deposits a subaqueous RMB offshore. (b) The channel and the RMB prograde basinward. (c) The channel and RMB continue progradation until the RMB reaches a critical size and the bar stagnates. (d) After stagnation the bar aggrades to sea level and the flow bifurcates. (e) Once the bar has stagnated a backwater forms at the bar front that propagates upstream morphodynamically, increasing bed aggradation and overbank flow. (f) The avulsion location is the position on the levee where the maximum shear stress is exerted for the longest duration. In each image the shoreline is located a few centimeters landward of the delta margin.
2.3.2 Propagation of morphodynamic backwater

After RMB stagnation the bar is an obstruction that creates a local bow wave or backwater with decreased velocity near the bar, which causes bed aggradation immediately upstream of the RMB (Figure 2-5a, location 1, $t/t_a = 0.6$ to $1.25$). The aggradation immediately upstream of the RMB then creates a new local backwater even farther upstream that leads to local aggradation (Figure 2-5a, locations 2 and 3, $t/t_a = \sim0.75$ to $1.25$). This upstream propagating “morphodynamic backwater” [Hoyal and Sheets, in press] is a wave of bed aggradation and water surface rise that causes a

![Figure 2-5](image)

Figure 2-5 (a) Time evolution of the bed and water surface since the start of the experiment ($t$) relative to the initiation of the avulsion ($t_a$) for the 3 locations marked on the trace of the DL9 shoreline. After the RMB stagnates (square on x-axis), the bed begins to aggrade (marked by upside down triangle) at location 1 and that change in aggradation propagates upstream to locations 2 and 3. The avulsion (star on x-axis) is initiated at location 3 soon after the morphodynamic backwater reaches that location. (b) Time evolution of the cumulative sediment volume ($V$) deposited within the distributary channel relative to the sediment volume deposited from RMB stagnation to avulsion ($V_a$). Timing of RMB stagnation ($t_b$) relative to $t_a$ is defined as zero and therefore negative values represent times before RMB stagnation. The areal extent of the channel used to calculate $V$ and $V_a$ was held constant for all experiments.
statistically significant increase (95% confidence level) in the net aggradation of the distributary channel network. In all the experiments the average net aggradation within the channel prior to RMB stagnation (Figure 2-5b, \( t_b/t_a = -0.5 \) to 0) is small. After RMB stagnation, the average net aggradation increases sharply to \(~0.4\) (Figure 2-5b, \( t_b/t_a = 0 \) to 0.5) due to the upstream-propagating morphodynamic backwater.

### 2.3.3 Morphodynamic backwater causes increased overbank flow

As the morphodynamic backwater moves upstream, the channel bed aggrades, the water surface rises, and there is increased flow over the levees. In five experiments (DL 2, 4, 5, 9, 12) there is sufficient temporal resolution to resolve the change in overbank flow through time. In those experiments, initially the percentage of wetted levee remains relatively constant (Figure 2-6a, \( t/t_a = 0 \) to 0.7). After the RMB stagnates, the morphodynamic backwater moves upstream and the percentage of wetted levee increases significantly (Figure 2-6a, \( t/t_a = ~0.7 \) to 1) until an avulsion is initiated. After an avulsion is initiated the percentage of wetted levee length begins to decline (Figure 2-6a, \( t/t_a > 1 \)) as the water surface elevation in the moribund channel decreases.
The cross levee flow generated by bed aggradation during the morphodynamic backwater is a necessary condition for avulsion because avulsions are initiated only after its passage (Table 2-1). For example, in DL5 (Figure 2-6b), the levee and water surface elevation at the avulsion site remain constant until the morphodynamic backwater passes and the flow depth over the levee crest increases. The increased flow...
depth increases the bed shear stress and the levee begins eroding, leading to avulsion initiation. Avulsion initiation is defined as the point in time when the levee at the avulsion site begins to experience runaway erosion (Figure 2-6b, $t/t_a = 1$). Over the entire delta the amount of cross-levee flow is not spatially uniform but depends upon levee heights and channel bed topography. Deep scour holes attenuate the aggradation signal of the morphodynamic backwater and keep the flow in bank, whereas shallower sections experience more aggradation relative to flow depth and consequently show more overbank flow.

<table>
<thead>
<tr>
<th>Lobe Number</th>
<th>Backwater reaches avulsion location (min)</th>
<th>Avulsion Occurs (min)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>341</td>
<td>410</td>
</tr>
<tr>
<td>4</td>
<td>197</td>
<td>382</td>
</tr>
<tr>
<td>5</td>
<td>151</td>
<td>225</td>
</tr>
<tr>
<td>9</td>
<td>375</td>
<td>512</td>
</tr>
<tr>
<td>12</td>
<td>218</td>
<td>320</td>
</tr>
</tbody>
</table>

Table 2-1: In the five experiments with sufficient temporal resolution, the avulsion always occurs after the morphodynamic backwater propagates upstream and passes the eventual avulsion site. This suggests that the passage of the morphodynamic backwater is a necessary condition for avulsion. Time is measured from river mouth bar stagnation.

2.3.4 Predicting avulsion location

The location of the avulsion depends not only upon the magnitude of the shear stress on the levee crest, but also upon the duration of its application. We propose that avulsion location is governed by time-dependent processes, rather than a strict threshold, because there must be a large enough shear stress on a levee to begin erosion, and it must be exerted long enough to erode a crevasse and construct an avulsion.
channel across the floodplain. This time-dependency suggests that the avulsion should occur where the cross-levee impulse per unit area of the flow, \( I \) \( (\text{kg m}^{-1} \text{s}^{-1}) \), is maximized:

\[
I_i = \int_0^{\Delta T} \tau_i \cdot dt
\]

for all \( \tau_i > 0 \), where \( \tau_i \) \( (\text{N m}^{-2}) \) is the bed shear stress at location \( i \) along the levee crest and \( \Delta T \) \( (\text{sec}) \) is total duration of cross-levee flow during the interval from RMB stagnation until just prior to avulsion initiation. Assuming steady, uniform flow, \( \tau \) equals \( \rho ghS \), where \( h \) is the water depth of flow crossing the levee crest and \( S \) is the floodplain slope measured from the levee crest to the shoreline along the path of steepest decent. The variable \( I \) is a proxy of the potential total amount of sediment transported during \( \Delta T \). A plot of \( I \) at various points along the levee crest during \( \Delta T \) (Figure 2-7A and B) indicates that the avulsion occurs at the location of maximum \( I \) with 93% accuracy. Within \( I \) is the product of two measured quantities (\( h \) and \( S \)) that have an associated error. We estimated the error propagation using the standard method of \( \text{Taylor} \) [1997]. When the error of each point in figure 2-7 is estimated the general interpretation does not change (Figure B1-B5, Appendix B). In general, the location of \( I_{\text{max}} \) lies a characteristic distance upstream from the RMB between 5 to 13 channel widths \( (n = 8) \), which is consistent with length scales of intra-delta lobe avulsions in the Mississippi delta \( \text{[Coleman and Gagliano, 1964]} \). These sites are far enough upstream so that the cross-levee slope is appreciable, but are far enough downstream so that flooding due to the morphodynamic backwater effect has been applied for a long time. Finding the location of \( I_{\text{max}} \) can be used to predict the avulsion location because in all
the experiments $I_{\text{max}}$ stabilizes near the avulsion site well before the avulsion occurs (Figure 2-8).

Figure 2-7  (A) The avulsion occurs in space where $I$, the cross-levee impulse, is greatest (marked by larger dot). $I_{\text{max}}$ corresponds to the location where the greatest average shear stress has been applied for the longest time. Each data point represents the average shear stress on the levees for one channel cross-section and is averaged in time from RMB stagnation until one scan prior to the avulsion. The channel cross-sections are averaged over 1.5 cm swaths and are calculated every 1.5 cm. The x-axis is shifted so that zero marks the avulsion location for each experiment. (B) The avulsion location ($\ell_a$) can be predicted ($\ell_p$) with 93% accuracy by finding the location of $I_{\text{max}}$ (large dots in A). Avulsion distances are measured from RMB crest to the avulsion location and non-dimensionalized by the average channel width ($\bar{W}$) in time and space.
2.4 Application to real deltas

These results provide guidelines for predicting avulsion locations on a delta, and the area and rate of creation of new wetlands associated with each avulsion. Even though antecedent conditions—such as, irregular levee topography, previously channelized flow paths [Aslan, et al., 2005; Jerolmack and Paola, 2007], and spatial variation of accommodation space—can influence the location of the avulsion, $I_{\text{max}}$ is still a reliable predictor of avulsion location. For example, in two experiments (DL2 and 9) the location of $I_{\text{max}}$ is coincident with low average bed shear stresses on the

Figure 2-8  Location of $I_{\text{max}}$ moves upstream after the RMB stagnates (zero on x-axis). Location of $I_{\text{max}}$ settles near the position of the eventual avulsion and remains there for a long time before the avulsion occurs. The black dots represent points of measurement.
levee, making those locations ostensibly poor candidates for avulsion. But, the partially channelized conditions of the floodplain adjacent to those locations permitted removal of sediment transported overbank, which sustained overbank flow, thereby maximized $I$, and eventually facilitated the avulsion.

### 2.4.1 Prediction of avulsion location in real deltas

The location of $I_{\text{max}}$ can be predicted on real deltas by a number of techniques. A high-resolution topographic survey of a delta coupled with a morphodynamic numerical model would allow simulation of repeated flooding on the delta and thus determination of $I_{\text{max}}$. Or, satellite measurements of water depth and water surface elevation [e.g., Alsdorf, et al., 2007] collected during floods could be used to estimate the location of $I_{\text{max}}$. Predicting the location of $I_{\text{max}}$ would benefit wetland restoration strategies [Reed and Wilson, 2004] that rely on man-made crevasses (e.g. West Bay Sediment Diversion on the Mississippi delta) to create wetlands. Crevasses placed at locations of $I_{\text{max}}$ would take advantage of the natural avulsion process and would insure that man-made crevasses do not disrupt the equilibrium of the delta.
2.4.2 Prediction of wetland growth in real deltas

Once an avulsion is created at the location of $I_{\text{max}}$ it is possible to estimate the future area and rate of wetland creation. Here we define wetlands as the partially inundated area of the intra-delta lobe lying adjacent to the distributary channel and its levees (Figure 2-4d). In this study the intra-delta lobe length ($L$) ranges from 5 to 13 channel widths ($n = 8$) and the average lobe width ($\bar{W}$) ranges from 2 to 6 channel widths ($n = 8$). Because $\bar{W}$ of an intra-delta lobe is related to $L$ ($\bar{W} = 0.23L + 1.78$, $R^2 = 0.43$), lobe area ($L \cdot \bar{W}$) and therefore wetland area can be predicted from the lobe length. As noted earlier, lobe length is a function of the jet momentum flux at the river mouth, $M$, and grain size [Edmonds and Slingerland, 2007]. Thus the area of future wetlands depends upon the location of $I_{\text{max}}$ because discharge increases up-delta and therefore so does $M$.

The rate at which new wetland area is created depends upon the speed of the morphodynamic backwater compared to the rate of RMB construction. If the speed of the morphodynamic backwater is fast compared to RMB construction, then avulsions occur quickly and tile the near-shore shallow water with relict intra-delta lobes that evolve to wetlands. If the speed of the backwater is slow, then lobe construction will continue into deeper water where less sub-aerial land is created per unit time because of the increased accommodation space. Therefore, in two delta lobes with the same $M$ and $I_{\text{max}}$, the lobe with a faster morphodynamic backwater will produce more wetland area per unit time.
2.4.3 How does the avulsion process interact with other network forming processes?

The relative rates of the upstream propagation of the morphodynamic backwater and the construction of RMBs may also help elucidate controls on delta morphology. If the upstream propagation of the morphodynamic backwater is slow relative to RMB construction then the system will continue to prograde basinward and bifurcate via mouth bar deposition. If, on the other hand, the upstream propagation is fast compared to RMB construction then the avulsion process will dominate delta morphology.

2.5 Conclusions

Here we have used physical experiments to present a clear description of avulsion mechanics and prediction of avulsion location in river-dominated deltas. The results clearly demonstrate that a class of avulsions are controlled by downstream processes, such as a growing river mouth bar (RMB), that cause an upstream migrating wave of bed aggradation and overbank flow. The avulsion timing and location can be predicted by finding the levee location of the maximum shear stress for the maximum time. To insure that the process of avulsion and wetland creation remains active in real deltas, the dredging of distributary channels and RMBs should be minimized because it disrupts the bed aggradation from the upstream-propagating morphodynamic backwater.
The extent to which the processes described in this paper actively participate in causing full-scale delta avulsions, which occur many more channel widths upstream of the RMB, remains an open and interesting question. Furthermore, the conceptualization of avulsion as a time dependent process, rather than a threshold process [e.g., Mackey and Bridge, 1995; Jones and Schumm, 1999; Jerolmack and Paola, 2007], may improve the prediction of avulsion location in other fluvial systems, such as alluvial fans, and anabranching and meandering rivers.
Chapter 3: Stability of delta distributary networks and their bifurcations

Doug Edmonds*
Rudy Slingerland

The Pennsylvania State University, Department of Geosciences, 513 Deike Building, University Park, State College, PA 16802, United States

*Corresponding author

Published in Water Resources Research
Received 12 March 2008
revised 18 June 2008
accepted 23 June 2008
published 18 September 2008

Abstract

Delta distributary networks are created by bifurcating channels that commonly split their discharges unequally. The origin and stability of these asymmetrical fine-grained, cohesive bifurcations are investigated here using Delft3D, a morphodynamic flow and sediment transport model. Results are compared to bifurcations on the Mossy delta, Saskatchewan, Canada that have remained stable for decades. Over a range of channel aspect ratios, friction factors, and Shields Numbers, we find three equilibrium functions relating the discharge ratio of the bifurcate arms at equilibrium to Shields number. One function defines symmetrical configurations (equal partitioning of discharge), while the other two define asymmetrical configurations (unequal partitioning of discharge).

Discharge asymmetries and morphologies of Mossy delta bifurcations are consistent with these predictions. Among the equilibrium bifurcations, only the asymmetrical type is stable to perturbations such as a partial closing of one throat. This possibly explains why asymmetrical bifurcations are more common in nature.
3.1 Introduction

A survey of the world’s river-dominated delta networks reveals that distributary channels rarely split their water discharges equally as they bifurcate into multiple channels. Rather, the discharges, and consequently the channel widths, depths, and sediment loads, are usually asymmetrical, seemingly representing a stable configuration [Edmonds and Slingerland, 2007] (Figure 3-1). This regularity seems surprising given the complexity of distributary channel mechanics, and suggests that morphodynamic feedbacks are at work acting to stabilize the delta channel network. Because it is the bifurcations that create the network, we focus on them here and ask these questions: 1) can asymmetrical bifurcations be in equilibrium such that the hydraulic properties of

Figure 3-1 Two examples of distributary deltas with bifurcating channel networks. On average the fluvial channel bifurcations—wherein one channel splits into two—are asymmetrical; their bifurcate discharges are unequal. This is true for the bifurcations on the coastline and the more mature bifurcations farther up the delta. A) Composite aerial photograph of Mossy Delta, Saskatchewan, Canada from 2003. Individual photos are from Information Services Corporation [2003]. The white line on the east side of the delta is the shoreline. The numbers mark the locations of the eight bifurcations in this study. B) Image of Wax Lake Outlet delta, LA from 1998 USGS aerial photography. Inset maps shows delta locations marked by a circle.
each bifurcate channel are adjusted to just transport the water and sediment given to it?

2) If so, are they stable, equilibrium configurations that return to their equilibrium configuration when perturbed? 3) Will perturbations such as climate change [e.g., Ericson, et al., 2006; Lesack and Marsh, 2007] and upstream impoundment of sediment by dams [Syvitski, et al., 2005a], lead to increasing instabilities and degradation of these channel network and their bifurcations?

These questions are important because the distributary networks that bifurcations create provide valuable maritime infrastructure and fertile floodplains to humans, and a nutrient-rich habitat for a diverse and biologically valuable ecosystem [Olson and Dinerstein, 1998]. Here we attempt to answer the first two questions using a numerical model and field data to elucidate how asymmetrical deltaic bifurcations function and under what conditions they are stable. Bifurcations are stabilized by processes operating locally, such as division of sediment at the bifurcation point, and processes operating globally, such as delta-scale changes in water surface slope. We restrict our analysis of distributary network stability to those processes acting locally on the bifurcations.

3.2 Present understanding of fluvial channel bifurcations

3.2.1 General characteristics of bifurcations

Among the occurrences of channel bifurcations we make the distinction between coarse-grained systems and fine-grained non-cohesive and cohesive systems. The
former seem to adjust their hydraulic geometry to maintain a Shields stress ($\Theta$) at about 1.4 times the critical Shields stress [Parker, 1978], whereas the latter maintain a $\Theta$ of about 1 for mixed load channels and about 10 for suspended load channels [Dade and Friend, 1998]. $\Theta$ is defined as

$$\Theta = \frac{\tau_o}{(\rho_s - \rho)gD_{50}}$$

(3-1)

where $\tau_o$ is the fluid shear stress (N/m$^2$), $\rho_s$ is the sediment density (kg/m$^3$), $\rho$ is the water density (kg/m$^3$), $g$ is acceleration due to gravity (m/s$^2$), and $D_{50}$ is the median bed grain size (m).

Even though coarse-grained (hereafter termed low $\Theta$) and fine-grained (hereafter termed high $\Theta$) bifurcations are thought to arise from different processes, ranging from flow splitting around bars to avulsion, they exhibit intriguingly organized and similar behaviors. On average, fluvial channel bifurcations are asymmetrical. Edmonds and Slingerland [2007] measured widths of the bifurcate channels on the world’s distributary deltas and found that the width ratios cluster around 1.7:1 ($n = 160$). A similar comprehensive study is missing for braided streams, but limited observations suggest that bifurcate width ratios for braided streams cluster around 1.5:1 ($n = 8$) [Zolezzi, et al., 2006]. For both cases, limited data indicate that channel widths are hydraulically adjusted to the discharge, and therefore the depths and discharges probably also are asymmetrical. The equilibrium configuration of asymmetrical bifurcations and their degree of stability are open and interesting questions that we will address in this study.
3.2.2 Low \( \Theta \) bifurcations

By far the most studied bifurcation type has been within low \( \Theta \) channels, whether by field observation [Davoren and Mosley, 1986; Ashmore, et al., 1992; Ashworth, et al., 1992; Bridge, 1993; Ashworth, 1996; Richardson and Thorne, 2001; Zolezzi, et al., 2006; Frings and Kleinhans, 2008], flume studies [Federici and Paola, 2003; Zanichelli, et al., 2004; Bertoldi and Tubino, 2005; Islam, et al., 2006; Bertoldi and Tubino, 2007], or by numerical modeling [Repetto, et al., 2002; Bolla Pittaluga, et al., 2003; Dargahi, 2004; Zanichelli, et al., 2004; Hall, 2005; Wu and Yeh, 2005; Kleinhans, et al., 2006; Miori, et al., 2006].

Present understanding of low \( \Theta \) bifurcation stability is summarized in Miori et al. [2006] and Bertoldi and Tubino [2007] who built on the pioneering approach of Bolla Pittaluga et al. [2003]. Bolla Pittaluga et al. [2003] approached the problem with a 1D numerical model of steady, uniform flow through a bifurcation. They discovered that the bed ramp, defined as the topographic rise in elevation from the unbifurcated reach to the shallower of the two downstream channels, steered different amounts of bedload to each downstream channel enabling an asymmetrical stable, equilibrium solution. Their model predicts that as \( \Theta \) increases in the unbifurcated reach, the stable, equilibrium bifurcate discharge ratio (larger channel: smaller channel) should decrease. Miori et al. [2006] improved on the Bolla Pittaluga model by allowing channel width to vary according to hydraulic geometry rules. They also produced asymmetrical stable, equilibrium bifurcations, and found that the final stable function depends on whether a bifurcation forms through incision of a new channel, or flow splitting around a mid-
channel bar. Other work has shown that stable solutions can be a function of an upstream meander bend [Kleinhans, et al., 2006] or the bifurcation angle magnitude [Mosselman, et al., 1995].

3.2.3 High $\Theta$ bifurcations

There has been much less research on high $\Theta$ bifurcations. Only a few field studies exist [Axelsson, 1967; Andren, 1994; Sloff, et al., 2003; Edmonds and Slingerland, 2007], experimental studies are hampered by scaling considerations [Zanichelli et al., 2004], and theoretical studies are limited to Wang et al. [1995], and Slingerland and Smith [1998].

Development of an adequate stability theory for high $\Theta$ bifurcations also lags behind the low $\Theta$ case. It may be that the theory for low $\Theta$ bifurcations also applies to high $\Theta$ bifurcations; however, this idea is untested and hinges on what roles the suspended load and sediment cohesiveness play. Wang et al. [1995] considered a bifurcation where the two bifurcate channels flow into a lake. These authors introduce an empirical nodal point boundary condition that controls the partitioning of water and sediment into the downstream branches. Their 1-D, steady, uniform flow analysis shows that the system contains only one stable state: a symmetrical division of discharge with both branches open. Slingerland and Smith [1998] improved on Wang et al.’s model by using the 1-D St. Venant equations coupled with suspended sediment and Exner’s equation for the case of river avulsions. They showed that symmetrical
configurations are unstable to small perturbations. However, their analyses focused on conditions for avulsion and not exactly on bifurcation stability.

In summary, our understanding of high $\Theta$ bifurcations is severely limited. The current theoretical treatments are oversimplified and do not consider nonuniform flows that are known to be important in some bifurcations [Dargahi, 2004]. Current numerical treatments also rely on an artificial internal boundary condition, or nodal point relation, to distribute sediment at the bifurcation. A more sophisticated modeling approach that accounts for the effect of unsteady, nonuniform flow, and allows the system to develop its own nodal point relation is needed. Perhaps most importantly, field data are needed to validate the stability studies of high and low $\Theta$ bifurcations.

Given that high $\Theta$ bifurcations are the dominant type on navigable rivers, there is a pressing need for detailed field data and improved theoretical modeling. To this end, our approach is to use numerical modeling to define the equilibrium solutions for high $\Theta$ bifurcations and then perturb those configurations to see if they are stable. We then use field data to validate the predictions. Important objectives are: (1) to define the stability functions for high $\Theta$ bifurcations and compare them to low $\Theta$ bifurcations; (2) to define the hydraulic and sedimentary processes that create stable, asymmetrical, high $\Theta$ channel bifurcations; and (3) to understand why bifurcating channels are generally asymmetrical with respect to their discharges, widths, and depths.
3.3 Numerical model description

3.3.1 Model description

We model the processes within a fluvial-channel bifurcation using the computational fluid dynamics package Delft3D. Delft3D simulates fluid flow, waves, sediment transport, and morphological changes at time scales from seconds to years and has been validated for a wide range of hydrodynamic, sediment transport, and scour and deposition applications in rivers, estuaries, and tidal basins [Hibma, et al., 2004; Lesser, et al., 2004; Marciano, et al., 2005; van Maren, 2005]. The equations of fluid and sediment transport and deposition are discretized on a curvilinear finite difference grid and solved by an alternating direction implicit scheme. An advantage of Delft3D is that the hydrodynamic and morphodynamic modules are fully coupled; the flow field adjusts in real time as the bed topography changes.

3.3.2 Governing equations

Delft3D solves the three-dimensional non-uniform, unsteady, incompressible fluid flow Reynolds equations under the shallow water and Boussinesq assumptions. The equations consist of conservation of momentum, conservation of mass, and the transport equation. The vertical eddy viscosities are defined using a $\kappa-\varepsilon$ turbulence closure scheme and the horizontal eddy viscosities are defined using a horizontal large eddy simulation that relates the horizontal fluid shear stress to the horizontal flow velocities. We did numerical experiments with and without the horizontal large eddy simulation and found it did not have an appreciable effect on the final solutions.
Therefore to reduce computational time, the horizontal large eddy simulation was not used and the horizontal fluid eddy diffusivities in all experiments are set to a constant value of 0.0001 m$^2$/s in the $x$ and $y$ directions. All results presented here use the vertically integrated 2D equation set in Delft3D because the equilibrium solutions vary little from the 3D solutions (5 equally sized computational layers in the vertical) by only a maximum of 15 percent of the equilibrium discharge ratio.

Delft3D has separate mathematical treatments for the erosion and deposition of cohesive and noncohesive sediment. Cohesive sediment is defined as silt-sized and finer, whereas noncohesive sediment is defined as sand-sized and coarser. The formulation for cohesive sediment erosion and deposition is based on work by Partheniades [1965] and Krone [1962], whereas the formulation for noncohesive sediment erosion and deposition is based on the Shields curve.

Cohesive and noncohesive sediment can be transported as bedload or suspended load depending on the grain size and the flow strength. Bedload transport rate per unit width is calculated from van Rijn [1984]. The magnitude and direction of the bedload transport vector is adjusted for favorable and adverse longitudinal slopes according to Bagnold [1966] and for transverse slopes according to Ikeda [1982]. Suspended load transport rate is calculated by solving the vertically-integrated three-dimensional diffusion-advection equation, where the sediment eddy diffusivities are a function of the fluid eddy diffusivities. Gradients in the sediment transport vectors are used to determine changes in bed topography using the Exner equation. For a more detailed discussion on the mathematics of Delft3D and the flow/topography interactions see Lesser et al. [2004].
3.4 Numerical modeling approach

3.4.1 Model grid considerations

For our experiments, we designed a computational grid with a straight unbifurcated (upstream) reach and two bifurcate reaches (Figure 3-2A). The unbifurcated reach is defined as channel \( a \), the bifurcate channel with the smaller discharge is channel \( b \), and the bifurcate channel with the larger discharge \( c \).

The grid is perfectly symmetrical about the centerline of the unbifurcated reach with a bifurcation angle between the two bifurcate channels of 55°. The channels have fixed walls and the top-width of each bifurcate is approximately one-half the width of the unbifurcated channel.

Each bifurcate channel has a nondimensional length \( (L') \) of approximately 12.5 which is consistent with the average \( L' \) of approximately 14 reported by Edmonds et al. [2004] from a survey of 24 distributary deltas throughout the world. \( L' \) is defined as \( L'=L/W \), where \( L \) is the dimensional channel length and \( W \) is the dimensional channel width. We did additional experiments with numerical grids that have longer bifurcate channels \( (L' = 37.5) \) to see if \( L' \) influenced our results. Equilibrium solutions for the longer bifurcate channels fall within 6 to 10% of the equilibrium solution for shorter bifurcate channels, leading us to conclude that the results presented here are insensitive to the range of \( L' \) measured by Edmonds et al. [2004]. In all cases our channel lengths are shorter than the backwater length scale (~15 kilometers for these experiments) as would be expected for distributary channels near the coastline.
Numerical results should also be independent of grid cell size [Hardy, et al., 2003]. We tested for grid independence and found that the results of this study were relatively insensitive to grid size. Therefore, we chose a grid that is numerically efficient yet still resolves topographic details in the evolving system. Each grid cell is a rectangle that is approximately two meters wide and 15 meters long with the long axis of the rectangle parallel to the flow direction. The time step in our experiments obeys the Courant-Frederichs-Levy criterion, and therefore the smallest cell determines the size of the maximum time step.

Grids in Delft3D should be smooth and each cell should be orthogonal in order to conserve mass and momentum. To achieve orthogonality around the bifurcation point (Figure 3-2B 2B), extra grid cells were added and an orthogonal transformation...
was applied using the Delft3D gridding software. The computational grid used in these experiments has a maximum deviation from orthogonality of 20 degrees. This orthogonality does not affect the solution; experiments with higher orthogonality achieved results similar to results with lower orthogonality.

3.4.2 Model setup and boundary and initial conditions

The variables thought to govern the behavior of bifurcations can be grouped into three dimensionless parameters: Shields number of the unbifurcated reach, $\Theta_a$, aspect ratio of the unbifurcated reach, $\alpha_a$, and friction factor of the system, $C'$, where

$$\alpha = \frac{W}{D}$$  \hspace{1cm} (3-2)

$W$ is the width (m) and $D$ is the depth (m) and

$$C' = \frac{C}{\sqrt{g}}$$  \hspace{1cm} (3-3)

$C$ is the dimensional Chezy roughness (m$^{1/2}$/s), and $g$ is the acceleration due to gravity (m/s$^2$).

The numerical modeling experiments use a range of these parameters to accurately represent fine-grained, cohesive fluvial bifurcations. The experiments are 2D vertically-integrated with one inlet and two outlets. The inlet boundary condition of channel $a$ is a steady, uniform discharge across the channel carrying an equilibrium sediment concentration. The outlet boundary conditions are steady, uniform free water surface elevations for channel $b$ ($h_b$) and for channel $c$ ($h_c$). The bed elevations at the downstream boundaries are allowed to adjust during the simulations.
At the inlet we prescribe equilibrium sediment concentrations that consist of a cohesive fraction of mud and a noncohesive fraction of fine-grained sand. Initially in the erodible substrate there is an equal proportion of evenly mixed noncohesive and cohesive sediment. We used a temporally and spatially invariant nondimensional Chezy roughness ($C'$) value of 12.5 and the aspect ratio ($\alpha$) of channel $a$ of approximately 16 for all runs. $\Theta_a$ varied from approximately 0.047 to 30.

The initial river bed topography for each numerical experiment consists of a uniform bed elevation in each channel where the initial bed elevation in channel $b$ is always higher than $a$ and $c$ (Figure 3-3). If there is a vertical offset between channel $a$ and $b$ (i.e. if there is a vertical step at the entrance to channel $b$) the model will not find an equilibrium solution because the local water surface slope induced by the offset causes channel $c$ to capture all the flow. However, if there is no vertical offset and the entrance is sufficiently smooth the model is not sensitive to initial conditions. To generate a smooth entrance we construct a bed ramp by linearly

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{figure3-3.png}
\caption{Two different initial bed ramp configurations tested in this study. The results of this study are insensitive to the initial bed ramp configuration therefore the top configuration was used. The location of the close-up is given in Figure 3-2A.}
\end{figure}
interpolating the bed elevation from channel $b$ approximately one or two multiples of $W_a$ upstream [Bolla Pittaluga, et al., 2003]. This initial condition permits a variety of different bed ramp configurations that will evolve to a single equilibrium solution (Figure 3-3). The experiments in this study used the bed ramp configuration in Figure 3-3 that extends approximately 2 channel widths upstream from the bifurcation point.

### 3.4.3 Obtaining an equilibrium bifurcation configuration

What is the appropriate metric for determining if a deltaic bifurcation is at equilibrium? Equilibrium deltaic systems are net depositional because the downstream boundary is changing due to delta progradation or changing sea level. However, as a first approximation we argue that we can assume that equilibrium deltaic bifurcations are adjusted for sediment bypass because the time scale for channel adjustment is very small compared to river mouth bar construction and channel building on deltas (~15 years for the Mossy delta). Therefore, we follow the definition of Miori et al., [2006]; bifurcations are in equilibrium if they do not change in morphology over some multiple of the morphological time scale ($T_m$), where

$$T_m = \frac{W_a D_a}{q_{sa}}$$

(3-4)

$W_a$ is width (m) in channel $a$, $D_a$ is the depth (m) in channel $a$, and $q_{sa}$ is the sediment transport rate per unit width ($m^2/s$) in channel $a$. $T_m$ is the duration over which the amount of sediment needed to fill a cross-section is transported through that cross-section. We consider a bifurcation to be in equilibrium if there is suspended and bedload transport in all reaches and the change in discharge ratio, $Q_r$, with time varies
by no more than 1% around the equilibrium value for at least 15 multiples of nondimensional time ($T_{ND}$), where

$$Q_c = \frac{Q_c}{Q_b}, \quad \text{and}$$

$$T_{ND} = \frac{T}{T_m}$$

(3-5)

(3-6)

$Q_c$ and $Q_b$ are the water discharges ($\text{m}^3/\text{s}$) in the channels with the larger and smaller water discharges, respectively, $T$ is the total time elapsed (days), and $T_{ND}$ is the nondimensional time, or the multiples of the morphological time scale elapsed during the computation.

To find equilibrium configurations in Delft3D we start with generic bed bathymetry with $D_r \neq 1$, where $D_r$ is the initial average depth ratio of the bifurcate channels, and a generic bed ramp between the channel $b$ and channel $a$. We then adjust the Shields stress in channel $a$ ($\Theta_a$) until we find the value that produces an equilibrium bifurcation configuration from that set of boundary conditions and initial $D_r$ (Figure 3-4) if one exists. We use this method to find the equilibrium solution because the nonlinear nature of the equations demands that the initial and boundary conditions be close to the

![Figure 3-4 Evolution of the discharge ratio for different initial $\Theta_a$. The initial conditions are generic bed and water surface topography at $T_{ND}$ equals 0. The bold line is an invariant $Q_r$ over many $T_{ND}$ and is considered to be an equilibrium solution for that set of boundary conditions. Variables defined in text and nomenclature list.](image)
solution for the model to recover that solution. As such, if the initial $\Theta$, in Figure 3-4 is much larger than the equilibrium value, $Q_r$ goes to 1, and if the initial $\Theta$ is much smaller than the equilibrium value, $Q_r$ goes to infinity, i.e., one channel closes completely. The bold line in Figure 3-4 represents a single equilibrium bifurcation solution. To build an entire equilibrium diagram we chose a number of different initial $D_r$ values and found the corresponding $\Theta$, that resulted in an equilibrium bifurcation.

Delft3D allows the user to speed up the bed adjustments by multiplying the deposition or erosion rate in each time step by a morphological scale factor. A series of sensitivity experiments showed that the final solution is insensitive to a morphological scale factor less than 250. We used a factor of 50. Approximately 100 simulations were conducted to define an equilibrium field.

3.5 Results

There are two classes of equilibrium bifurcations, those which have equal water surface elevations at their downstream boundaries, such as bifurcations on deltaic coasts (Figure 3-1), and those which have an imposed advantage due to different water surface slopes, such as more mature bifurcations further up delta. The differences between the equilibrium states of each class are not well defined. To this end, we conducted two sets of equilibrium experiments, one with equal water surface elevations at the outlets of channels $b$ and $c$ ($h_b = h_c$) and one with an imposed water surface slope advantage from unequal water surface elevations at the outlets of channels $b$ and $c$ ($h_b \neq h_c$). Delft3D
finds symmetrical \((D_r, Q_r = 1)\) and asymmetrical \((D_r, Q_r \neq 1)\) equilibrium functions for both experiment sets. We first present a description of the typical equilibrium bifurcation with a symmetrical and asymmetrical configuration common to both experiments. Then we summarize the results from all experiments in a bifurcation equilibrium diagram and comment on the stability of the equilibrium solutions.

3.5.1 Description of the General Bifurcation Equilibrium Configuration

The equilibrium bifurcations created in this study share the same basic topographic and hydraulic forms and features: 1) across the entrances to channels \(b\) and \(c\) there is cross-channel variation of water surface elevation and bed topography (Figure 3-5); 2) there is a positive bed ramp at the entrance to channel \(b\) and a negative bed ramp at the entrance to channel \(c\) (Figure 3-5); 3) the thalwegs for channels \(b\) and \(c\) are located on their southern and northern banks, respectively (Figure 3-5); and 4) in equilibrium asymmetrical configurations, the water surface topography is complex around the bifurcation point (Figure 3-6).
Figure 3-5 Examples of symmetrical (A) and asymmetrical (B) water surface elevation and bed topography at equilibrium as computed by Delft3D. The solutions have the following characteristics: 1) at the entrances to channels \( b \) and \( c \) there are regions of the water surface that are elevated (a) and depressed (b) and topographic features of positive (c) and negative (d) relief; 2) there is a positive bed ramp (e) at the entrance to channel \( b \) and a negative bed ramp (f) at the entrance to channel \( c \); and 3) the channel thalwegs (g) of channels \( b \) and \( c \) are located along the southern and northern banks, respectively. The bed topography is measured with respect to the downstream water surface elevation boundary, which is zero in these cases. The white dotted lines in (B) refer to locations of water surface elevation profiles in Figure 3-6. In this example \( Q_a = 257.5 \), \( Q_b = 75 \), and \( Q_c = 182.5 \).
The cross channel variations in water surface elevation and bed topography at the entrances to channels \( b \) and \( c \) are the result of the interaction of the flow with an obstruction (i.e., the point of the bifurcation and the bed ramp). This can be understood by considering a streamline through the middle of channel \( a \) that intersects the bifurcation point of a symmetrical bifurcation \( (D_r, Q_r = 1) \). If energy along that streamline is conserved, the water surface must rise because kinetic energy is converted to potential energy at the bifurcation point where the velocity goes to zero. In our experiments we observe a rise in water surface of 0.5 to 1 centimeter (a in Figure 3-5), which is similar to the 1 to 2 centimeter rise predicted by the Bernoulli equation. Additionally, streamlines just north and south of the bifurcation point respond similarly to the bifurcate channel curvature and the water surface is also elevated (a in Figure 3-5A). The elevated water surface around the bifurcation point creates a low velocity zone at the bifurcation point and in turn sediment is deposited in the entrances to channels \( b \) and \( c \) (c in Figure 3-5A). If at the entrances to channels \( b \) and \( c \) part of the water surface is elevated and the corresponding velocity is low, to conserve mass through the entire cross section, part of the water surface must also be depressed (b in Figure 3-5A) and the local velocity increased on the outside bank. The higher velocity produces scour holes (d in Figure 3-5A) on the northern and southern banks at the entrances to channels \( b \) and \( c \), respectively.

Now consider a streamline that intersects the bifurcation point in an asymmetrical bifurcation \( (D_r, Q_r \neq 1) \) (Figure 3-5B). The same general water surface and topographic forms described above are observed. However, the bed ramp is more effective at increasing the water surface elevation because in addition to
conversion of kinetic to potential energy due to the obstruction of the bed ramp, the water surface is also elevated due to the increasing elevation of the bed. The presence of the bed ramp makes the elevated water surface asymmetrical in the entrances to channels \( b \) and \( c \) (a in Figure 3-5B). This also creates asymmetrical bed topography (d in Figure 3-5B) that is a skewed version of the symmetrical case. At equilibrium the bed ramps of both channels extend \( 0.9W_a \) to \( 2.25W_a \) upstream from the bifurcation point at low and high equilibrium values of \( Q_r \), respectively (e and f in Figure 3-5A, B).

The bifurcate thalwegs are located on the northern and southern banks of channels \( b \) and \( c \) respectively (g in Figure 3-5A,B), because upstream of the bifurcation fluid parcels in channel \( a \) have a momentum vector oriented parallel to the banks of channel \( a \). As those fluid parcels enter channels \( b \) and \( c \) they are not immediately aligned with the banks of channels \( b \) and \( c \). The inherited momentum orientation from channel \( a \) forces the high velocity thread to the southern and northern banks of channels \( b \) and \( c \), respectively. In turn, this causes the thalweg downstream of the bifurcation point to be located in the same position (g in Figure 3-5A, B).

In an equilibrium asymmetrical bifurcation the water surface elevation profiles down the middle of channels \( b \) and \( c \) are nonuniform; the water surface at the entrance to channel \( b \) is elevated, while at channel \( c \) it is depressed relative to a uniform water surface slope (Figure 3-6). The nonuniformity of the water surface extends upstream \( 2.5W_a \) to \( 5W_a \) at low and high equilibrium values of \( Q_r \), respectively. The cause of the nonuniform water surface configuration around the bifurcation is related to flow past an obstruction and the presence of a bed ramp in each channel.
The morphodynamic feedbacks among the dynamic water surface elevation, flow velocities, bed slopes, and sediment transport vectors create an equilibrium bifurcation in which $Q$ and sediment discharge ($Q_s$) of the bifurcate channels are delicately adjusted to just transport the sediment and water delivered to them. In this example (Figure 3-5 and Figure 3-6) the ratios are $Q_r$ and $Q_{sr} \approx 2.5$.

![Figure 3-6 Equilibrium water surface profiles for an asymmetric bifurcation. The profile of channel $b$ is elevated, while the profile of channel $c$ is depressed relative to a uniform slope (black dashed line). The elevation and depression extend well upstream of the bifurcation (black triangle). The locations of the profiles correspond to the while dotted lines in Figure 3-5.](image)

3.5.2 Equilibrium diagram

Similar to the theoretical analysis of coarse-grained bifurcations [Bolla Pittaluga, et al., 2003; Miori, et al., 2006], our equilibrium diagram for fine-grained, cohesive bifurcations is characterized by three equilibrium functions in $\Theta_a$ space: 1) at all values of $\Theta_a$ above transport and with $h_b = h_c$, there is an equilibrium function with $D_r$ and $Q_r = 1$ (Figure 3-7A); 2) at relatively low values of $\Theta_a$ and with $h_b = h_c$, there is an equilibrium function with $D_r$ and $Q_r \neq 1$ (Figure 3-7B); and 3) at relatively high
values of $\Theta_a$ and with $h_b \neq h_c$, there is an equilibrium function with $D_r$ and $Q_r \neq 1$ (Figure 3-7C). The symmetrical equilibrium function occurs through all values of $\Theta_a$ greater than the critical Shields stress, $\Theta_{crit}$ (Figure 3-7A). This is an unsurprising result and has also been found with other numerical models [Wang, et al., 1995; Slingerland and Smith, 1998].

More surprisingly, at $\Theta_a < 2.3$ and $h_b = h_c$ there are asymmetrical equilibrium bifurcations, whose equilibrium $Q_r$ is a positive function of $\Theta_a$ (Figure 3-7B). The function stops at $Q_r > 6.5$ because $Q_b$ becomes so small that $\Theta_b$ falls below $\Theta_{crit}$ for noncohesive sediment. Delft3D predicts that $Q_r$ increases (becomes more asymmetrical) as $\Theta_a$ increases, which is opposite to predictions for coarse-grained bifurcations determined using numerical solutions of the steady, uniform flow equations.
Why does Delft3D predict an increase in equilibrium $Q_r$ with increasing $\Theta_a$?

We computed the static flow and sediment transport fields at various values of $Q_r$ (or bed ramp height) while holding $\Theta_a$ constant. The bed was not allowed to deform. An increase in bed ramp height ($\eta$) causes an increase in $Q_r$ and sediment discharge ratio, $Q_{sr}$ ($\text{m}^3/\text{s}$), because more discharge and sediment are diverted to channel $c$ than to channel $b$ by the larger $\eta$. An increase in $\eta$ also increases the water surface slope in channel $b$ (Figure 3-8). A steeper water surface slope in channel $b$ requires more bedload to be delivered to the entrance of channel $b$ to remain in equilibrium, and this can only happen if $\Theta_a$ is larger. The result is a counterintuitive, inverse relationship between water and bedload discharge (Figure 3-9) in the smaller discharge channel of equilibrium.

Figure 3-8 For a given $\Theta_a$, an increase in bed ramp height ($\eta$) increases the water surface at the entrance to channel $b$. The water surface elevation is relative to $h_b$ and $h_c$ (which are equal).

Figure 3-9 In equilibrium bifurcations computed in this study there is an inverse relationship between bedload transport ($Q_{s\text{bedload}}$) in channel $b$ and discharge ($Q_b$) in channel $b$. See text for details.
bifurcations, because an increase in $Q_r$ simultaneously causes an increase in water surface slope (and hence bedload transport) and a decrease in discharge of that channel. Therefore the trend of this equilibrium function is a consequence of the elevated and depressed water surface (Figure 3-6 and 8) around the bifurcation point that allows for equilibrium asymmetrical energy slopes and sediment transport rates in the bifurcate arms.

At $\Theta_a > 2.1$ with $h_b \neq h_c$, Delft3D predicts that equilibrium $Q_r$ decreases as $\Theta_a$ increases for a given combination of $h_b$ and $h_c$ (Figure 3-7C). For this particular realization $h_c$ was set 2.5 millimeters lower than $h_b$, which makes the water surface slope in channel $c$ 10% steeper than channel $b$. The lowest $\Theta_a$ for which this equilibrium solution exists depends on the $h_c/h_b$ ratio. As $\Theta_a$ increases, the equilibrium $Q_r$ approaches 1 asymptotically. This occurs because the slope advantage imposed by different $h_b$ and $h_c$ becomes an increasingly small percentage of the water surface slope at high $\Theta_a$. Nonetheless it is interesting that the asymmetrical function with $h_b \neq h_c$ does not exist for all values of $\Theta_a$. This function terminates at low $\Theta_a$ because the water surface slope down the favored bifurcate arm is steep compared to the water surface slope in channel $a$, and all the sediment and discharge are routed down the favored bifurcate channel. Thus for each realization of $h_b \neq h_c$, there is a threshold point in $\Theta_a$ space below which the water surface slope ratio between the unbifurcated reach and the favored bifurcate reach is too large to maintain an equilibrium asymmetrical solution.
3.5.3 The effect of changing channel roughness \((C')\) and aspect ratio of upstream channel \((\alpha_a)\) on the equilibrium functions

It is important to remember that the results presented so far are for a particular combination of \(\alpha_a\) and \(C'\). Additional numerical experiments show that at larger (smaller) values of \(\alpha_a\) the relatively low \(\Theta_a\) equilibrium function (Figure 3-7B) shifts to a higher (lower) equilibrium \(Q_r\) for a given \(\Theta_b\). At larger (smaller) values of \(C'\) the relatively low \(\Theta_a\) equilibrium function (Figure 3-7B) shifts to a lower (higher) equilibrium \(Q_r\) for a given \(\Theta_b\). At larger (smaller) values of \(\alpha_a\) the relatively high \(\Theta_a\) equilibrium function (Figure 3-7C) shifts to a lower (higher) equilibrium \(Q_r\) for a given \(\Theta_b\). At larger (smaller) values of \(C'\) the relatively high \(\Theta_a\) equilibrium function (Figure 3-7C) shifts to a higher (lower) equilibrium \(Q_r\) for a given \(\Theta_b\).

3.5.4 Are these equilibrium configurations stable to perturbations?

To test if the equilibrium configurations are stable we added a perturbation to the equilibrium river bed topography and let the model continue to compute forward. If the perturbation was damped and the bifurcation returned to the original equilibrium form then that configuration is considered to be in a stable equilibrium. We follow previously published methodology [Bolla Pittaluga, et al., 2003] and add a small sediment bump (~ 50 cm high and 10500 cm³ or approximately a 5-10 percent net change in channel cross-sectional area) in the middle of channel \(b\).

All three equilibrium functions (Figure 3-7) are stable to small perturbations; when subjected to a perturbation in the shallower bifurcate channel the system returns
to the equilibrium configuration (Figure 3-10). Additional numerical experiments prove that all configurations in the three equilibrium functions are stable to a small sediment bump in the middle of channel \( b \). However, if the size of that sediment bump is sufficiently large (~100 cm high and 450000 cm\(^3\) or a 30-40 percent net change in channel cross-sectional area), the symmetrical configuration is no longer stable. This is true for all symmetrical solutions over the range of \( \Theta_a \) in Figure 3-7. For example, if the symmetrical equilibrium solution with \( Q_r = 1 \) and \( \Theta_a = 1 \) is perturbed with a large sediment bump the new equilibrium solution has \( Q_r = 1.1 \). These results suggest that asymmetrical bifurcations are more stable to perturbations compared to symmetrical bifurcations.
3.6 Validation of model results using field data

To validate the theoretical predictions (in the sense of Hardy, et al. [2003]) we collected river bed topography, hydraulic data, and water surface elevations on eight natural bifurcations in the Mossy delta, Saskatchewan, Canada (Figure 3-1A). River bed elevations were collected using an EAGLE FishElite 500c single-beam echo sounder. Channel water discharges were measured with an acoustic Doppler current profiler at near bankfull flow stage (Parsons and Best, pers. comm.). Water surface elevation data were collected by mounting a Leica differential global positioning system (dGPS) rover (receiver) in a boat floating down the middle of the channels recording water surface elevations every second. The elevation data were processed with Ski-Pro v. 3.0 using a base station of known elevation. This technique is advantageous because the high temporal resolution and vertical accuracy resolve the details of the water surface. To validate this technique, we floated the same river reach multiple times over different days and observed the features of water surface topography in each float.

The value of $\Theta_a$ for each bifurcation was calculated using measured bed grain sizes, water surface slopes, and channel geometries. Time series of channel geometry for most bifurcations indicate that their widths have adjusted in accord with hydraulic geometry scaling [Edmonds unpub. data], suggesting that Mossy delta bifurcations are in equilibrium with the flow field. Additionally, a serial aerial photographic analysis shows most bifurcations have been active for over 35 years [Oosterlaan and Meyers, 1995; Edmonds and Slingerland, 2007], and have not changed appreciably in planform. Thus, we take the Mossy delta bifurcations to be stable, equilibrium forms and if the
predictions of Delft3D are accurate these forms should compare favorably with the theoretical numerical modeling results.

The river beds and water surfaces of Mossy delta bifurcations are similar to the stable, equilibrium forms predicted in this study (cf. Figure 3-2 and 11). The river beds of the natural bifurcations have the same topographic features as the stable, equilibrium bifurcations produced in Delft3D (e.g., Figure 3-5). Additionally, the water surface profiles of Mossy delta bifurcations show elevated and depressed topographic configurations near the bifurcation similar to model predictions (Figure 3-12). The water surface is elevated at bifurcation points producing steeper water surface slopes in two-thirds of the lower discharge bifurcate channels ($n = 8$), indicating that the nonuniform water surface is a common feature in natural bifurcations.

The eight Mossy delta bifurcations generally plot on or near the equilibrium function in the stability diagram and exhibit a trend similar to theory (Figure 3-13). We take the favorable comparisons between predicted and observed bed topographies (Figure 3-11 and Appendix A, Figures A1-11), water surface topographies (Figure 3-12 and Appendix A, Figure A-12), and locations on the theoretical stability diagram (Figure 3-13) as support for the stability diagram.
Figure 3-11 River bed topography of natural bifurcations on the Mossy delta, Saskatchewan, Canada. The locations of Mossy delta bifurcations are marked on Figure 3-1. These data were collected at near bankfull flow stage in July 2006. The natural bifurcations have features similar to equilibrium bifurcations produced in Delft3D. There are depositional and scour features around the bifurcation point, there is a positive bed ramp from the main channel to the shallower channel, and the bifurcate thalwegs are located on the inner banks.
Figure 3-12 A) Water surface elevations on a bifurcation taken with a Leica dGPS. Bold line is a 50 meter running average. B) River bed elevations taken with a single beam echo sounder. C) Planview map of Mossy delta bifurcation number 5 (location in Figure 3-1) showing the locations of data track lines. Similar to Delft3D predictions, the channel with a lower $Q$ and a bed step (channel $b$) has an elevated water surface relative to the projected uniform water surface slope, while the channel with a higher $Q$ (channel $c$) has a depressed water surface.
3.7 Discussion

3.7.1 Comparison to previously published models

The asymmetrical bifurcation stability function presented here (Figure 3-7B) is different from previously published results [Bolla Pittaluga, et al., 2003; Miori, et al., 2006; Bertoldi and Tubino, 2007]. For braided river bifurcations the stable, equilibrium $Q_r$ decreases as $\Theta_a$ increases. We think our stability solution has the opposite trend because the fine-grained, cohesive system is sensitive to, and controlled by, the strongly nonuniform water surface topography at the bifurcation. Our results were computed for a high $\Theta$ system and it is unknown if the previously published stability solutions at low $\Theta$ will be as sensitive to the nonuniform water surface topography at the bifurcation.

Asymmetrical bifurcations are stable in braided rivers because the bed ramp significantly alters the sediment transport vector [Bolla Pittaluga, et al., 2003], thereby
allowing each bifurcate channel to receive different amounts of sediment due to the transverse sediment flux \( (Qs_T) \) at the bifurcation point. \( Qs_T \) is defined as

\[
Qs_T = Qs - Qs^*
\]

where \( Qs \) (m\(^3\)/s) is the sediment flux in a bifurcate channel for a given solution and \( Qs^* \) (m\(^3\)/s) is the sediment flux in that bifurcate channel for an symmetrical bifurcation with the same hydraulic conditions. The difference is the sediment that is being redirected due to the presence of the bed ramp. Our results from Delft3D confirm that the effect of transverse slopes on the sediment transport vector is a necessary condition for achieving asymmetrical stability, but interestingly for fine-grained cohesive bifurcations it is not a sufficient condition. If the transverse slope effect on bedload transport is removed in our models, \( Qs_T \) changes by only a few percent. Rather the dominant mechanism of sediment steering is topographic steering of flow and sediment trajectories due to the presence of the bed ramp and nonuniform water surface elevation at the entrances of the two bifurcate channels. This topography at the bifurcation can not be known \textit{a priori}, instead it is the result of the morphodynamical feedbacks between a co-evolving river bed and flow field.

3.7.2 Why are deltaic bifurcations asymmetrical?

As noted earlier, on average, deltaic bifurcations have asymmetrical widths, depths, and discharges in the bifurcate channels. Our results show that asymmetrical bifurcations are more stable to perturbations than symmetrical bifurcations. In addition, \textit{Bertoldi and Tubino} [2007] recently proposed a novel explanation for bifurcation
asymmetry that may also hold true for deltaic bifurcations. They noted that under super-resonant conditions, the presence of the bifurcation causes a transverse bed perturbation upstream that topographically steers more flow into one of the bifurcate channels.

The prevalence of asymmetrical bifurcations in nature implies that there must be perturbations that drive the bifurcation away from symmetry. The results from this study (section 5.4) show that symmetrical bifurcations are less stable; a large perturbation can force the symmetrical bifurcation to become asymmetrical. There are a variety of perturbations that could cause bifurcation asymmetry ranging from process perturbations, such as the bifurcation itself [Bertoldi and Tubino, 2007], alternating side bars [Miori, et al., 2007], and river meandering [Kleinhans, et al., 2006], to white noise perturbations, such as floods, circulation dynamics in the standing body of water, water surface slope advantages, and planform advantages. Therefore, we suggest that asymmetrical bifurcations are prevalent because a symmetrical bifurcation will eventually become asymmetrical due to myriad perturbations in nature. Once bifurcations are asymmetrical, the nonuniform water surface topography at the bifurcation and the effect of the bed ramp on the flow field provide feedbacks that keep asymmetrical bifurcations stable.
3.8 Conclusions

We have attempted to explain the origin of asymmetrical bifurcations of river channels by investigating their stability using a 2D vertically-integrated morphodynamic numerical model (Delft3D). The morphodynamic feedbacks between the evolving bed and water surface create three distinct equilibrium functions where the equilibrium discharge ratio ($Q_r$) is a function of the Shields stress in the unbifurcated reach ($\Theta_a$). The first function has a symmetrical division of discharge in the bifurcate channels; the other two are asymmetric. With equal downstream water surface elevations (no imposed advantage for either channel), the stable, equilibrium $Q_r$ becomes more asymmetrical as $\Theta_a$ increases because the water surface elevation at the bifurcation rises, steepening the water surface slope, and thereby requiring a higher $\Theta_a$ for equilibrium. For unequal downstream water surface elevations (imposed advantage for one channel) the equilibrium $Q_r$ becomes more symmetrical as $\Theta_a$ increases because at large $\Theta_a$ the water surface slope advantage imposed by unequal downstream boundaries is a small percentage of the overall water surface slope.

When subjected to a perturbation, such as a small sediment mound in a bifurcate channel, the asymmetrical bifurcations return to their equilibrium configuration whereas the symmetrical bifurcation moves to an asymmetrical stable equilibrium solution. Our results suggest that asymmetrical bifurcations are prevalent in nature because they are stable to a wider range of perturbations.

These results are supported by field data from bifurcations of the Mossy delta in Saskatchewan, Canada. Field hydraulic geometry data and a 60 year history of little change suggest that the Mossy delta bifurcations are in stable equilibrium with their
flow field. The Mossy delta bifurcations contain remarkably similar asymmetric bed geometries and water surface profiles to those predicted by Delft3D. Furthermore, when the Mossy delta bifurcations are plotted on the stability diagram they plot in stable space.
Chapter 4: The response of a delta distributary channel network to a perturbation

Doug Edmonds¹*
Rudy Slingerland¹
Jim Best²
John Bridge³
Dan Parsons⁴
Norm Smith⁵

¹The Pennsylvania State University, Department of Geosciences, 513 Deike Building, University Park, State College, PA 16802, United States

²University of Illinois, Departments of Geology and Geography and Vent Te Chow Hydrosystems Laboratory, 1301 W. Green St., Urbana, IL 61821, United States

³University of Binghamton, Department of Geological Sciences and Environmental Studies, PO Box 6000, Binghamton, NY 13902-6000, United States.

⁴University of Leeds, Institute of Geological Sciences, School of Earth and Environment, Woodhouse Lane, Leeds, LS29JT, United Kingdom

⁵University of Nebraska, Department of Geosciences, 214 Bessey Hall, PO Box 880340, Lincoln, NE 68588-0340, United States

*Corresponding author

To be submitted for publication to Sedimentology
Abstract

Delta distributary networks are intricate geomorphologic systems that create wetlands, valuable natural habitat for diverse species, and provide important infrastructure for industry. Given this, it would be useful to know if they are vulnerable to perturbations, especially since current studies predict changes to climate patterns and sea level, which are deltaic boundary conditions. We conducted a suite of numerical experiments based on the channel network of Mossy delta in Saskatchewan, Canada. In a series of ten experiments we assess the vulnerability of a channel network to perturbations by forcing the abandonment of various channels in the delta network and documenting the response. Results show that closing a channel reorganizes the water and sediment fluxes throughout the delta channel network. This occurs because the water surface elevation upstream must increase to generate enough force for the water and sediment to flow down the other bifurcate channel, which is undersized relative to the new higher, discharge. The increase in water surface elevation can travel far upstream and if other bifurcation points are nearby, additional discharge flows down those branches. This reorganization of water and sediment can affect the long-term evolution of a delta channel network. These results suggest that any assessments of the long term evolution of delta channel networks such as the Mississippi delta should determine the likelihood of channel closure from climatic perturbations.
4.1 Motivation

Rivers deltas are valuable natural environments that provide wetlands, biodiversity, industrial shipping lanes, and home to approximately 25% of the world’s population [Coleman, 1988; Olson and Dinerstein, 1998; Syvitski, et al., 2005b; Day Jr, et al., 2007]. Thus it is sobering that their position at the land-sea interface makes them especially vulnerable to perturbations generated as changes in global climate alter their base level [McCarthy, et al., 2001] and their hydrology [Trenberth, et al., 2007]. Dams on the World’s rivers are now sequestering 30% of the sediment that formerly fed them, and extraction of water and hydrocarbons under many deltas has resulted in two- to four-fold increases in subsidence relative to pre-anthropogenic rates [Syvitski, et al., 2005b; Syvitski, 2008].

At present, these perturbations are having a measurable and adverse effect on some deltaic environments [e.g. Lesack and Marsh, 2007], but it is not clear to what extent deltas will be affected. Studies to date have focused on how an entire delta system might adjust to large-scale climatic perturbations, such as changes in sea-level [Milliman and Broadus, 1989; Day, et al., 1995; Sanchez-Arcilla and Jimenez, 1997; Ericson, et al., 2006; Trenberth, et al., 2007]. Most of these studies are concerned with identifying which deltas are at risk. While this is an important first step in hazard mitigation, the next step is to predict the morphodynamic behavior of the whole delta distributary net and associated wetlands in response to specific perturbations.

Here we focus on how the distributary network of a river-dominated delta responds to closure of a single channel in the net. Focusing on the distributary network of a delta is a logical first step in understanding whole-delta response because it is the
network that feeds sediment and water to its various parts to maintain the delta in the face of coastal erosion and relative sea level rise. This study is limited to river-dominated deltas to simplify the problem. We assume that closure of one or a few distributaries is the simplest manifestation of the climatic and man-made perturbations discussed previously. Closure might occur, for example, due to increases in the frequency and occurrence of bank failure, destabilized alternate bars, or increased ice and wood jams. Numerical modeling experiments were performed with the purpose of understanding if closure of a channel has an effect on the entire delta network, if it can destabilize the network, and how the location of a closure within the delta network changes its effect.

4.2 Controls on river-dominated delta stability

Most of the previous work on delta stability has considered the long-term stability of the entire delta system. Stanley and Warne [1994] showed that Holocene deltas began accumulating 8500-6500 years ago as sea level rise decelerated. Deceleration of sea level rise also has been linked to the formation of distributary delta channel networks [Amorosi and Milli, 2001; Warne, et al., 2002]. In some cases acceleration of sea-level rise has altered the channel network pattern [Aslan and Autin, 1999]. Once a network is formed, its stability is a function of the stability of the individual bifurcations within the network. Bifurcations are formed when a delta progrades and its distributary channels split around river mouth bars deposited at the
delta margin, or when distributary channels avulse to new locations [Edmonds and Slingerland, 2007]. There has been a substantial amount of research on the stability of fluvial channel bifurcations in fine-grained [Wang, et al., 1995; Slingerland and Smith, 1998; Edmonds and Slingerland, 2008; Kleinhans, et al., 2008] and coarse-grained systems [Bolla Pittaluga, et al., 2003; Federici and Paola, 2003; Bertoldi and Tubino, 2005; Miori, et al., 2006; Zolezzi, et al., 2006; Bertoldi and Tubino, 2007]. We now know that fluvial bifurcations distribute water and sediment asymmetrically; one bifurcate arm takes roughly twice the water discharge and is usually deeper and wider than the other (for example, see the first bifurcation in the Mossy delta; Figure 4-1). These asymmetrical bifurcations are in equilibrium for a narrow range of Shields stresses for the upstream channel, and discharge ratios for the downstream channels. The asymmetrical configuration for both fine- and coarse-grained systems is stable; when each system is subjected to a small perturbation, such as a bank slump in a bifurcate channel, the asymmetrical configuration is recovered. Bifurcations are not stable to all perturbations because a curved channel segment in the unbifurcated reach can cause instability by changing the amount of water and sediment delivered to the bifurcate channels, and by inducing downstream bar migration that disrupts sediment partitioning in the throats of the bifurcate channels [Kleinhans, et al., 2008].

These conclusions on bifurcation stability apply to one bifurcation in isolation from the channel network that is subject to steady upstream and downstream boundary conditions. But in real delta networks the bifurcations are all connected. As one bifurcation adjusts to a perturbation there is the potential for the adjustment to ripple throughout the network. For instance, the presence of a bifurcation produces a
measurable backwater that occurs as water flows through a bifurcation [Edmonds and Slingerland, 2008]. This backwater can extend upstream far enough to affect the water surface slope of other bifurcates. Therefore, if a bifurcation (e.g., 2.1 in Figure 4-1) became unstable and one bifurcate channel was closed, the associated backwater would disappear and the water surface slope of the channel upstream would change (e.g., 1.1 in Figure 4-1). Depending on the magnitude of this change, it could change the stability of the bifurcations upstream and downstream, which in turn could disrupt other bifurcations, thereby propagating the original localized perturbation throughout the delta network. Therefore, even though it is true that single bifurcations are stable to perturbations, it is not necessarily true that a network of bifurcations is stable to perturbations, especially if the perturbation is large enough to close an entire channel.

How does a network of bifurcations respond to perturbations? Karssenberg and Bridge [2008] recently looked at this problem using a three-dimensional alluvial stratigraphy model that simulates distributive network development through avulsion. Even though they focused on network development upstream of the coastline and not explicitly within deltas, their numerical experiments serve as a useful guide for understanding how a delta channel network responds to perturbations. Karssenberg and Bridge observed how the character of the network changed under conditions of increased sediment supply, and rise and fall of sea-level over 10,000 years. During low rates of sea level rise the channel network was essentially unaffected and there was no discernable relationship between avulsion frequency and sea level rise. As the rate of sea level rise increased the avulsion frequency also increased. The avulsions began downstream and then moved upstream until they occurred over the length of the
domain. They observed that relatively fast rates of sea-level rise completely disrupted the stability of the original network. In another recent study, Hoyal and Sheets [in press] using physical tank experiments of deltas, suggested that perturbations, such as a river mouth bar growing at the delta tip, can trigger bifurcation instability. In some cases the instability of the bifurcation led to instability of the network. Although their data did not conclusively show that instability of the bifurcation directly caused instability of the network, it is suggestive.

In summary, there is evidence suggesting perturbations can rearrange a delta network configuration, but as yet there is no mechanistic predictive capability. Towards this end we investigate here the adjustments of a delta distributary network to closure of one channel segment at various locations within the network. To make the problem concrete the network is patterned after the Mossy delta network in Saskatchewan, Canada.
Figure 4-1  2003 Composite aerial photograph of Mossy Delta, Saskatchewan, Canada. Individual photos are from Information Services Corporation[2003]. A numerical grid constructed in Delft3D is overlain on the image. The bifurcations are numbered according to their position in the delta. The first number indicates the bifurcation order, where order refers to the number of times the flow along a channel has been split. The second number distinguishes among multiple bifurcations of the same order. Due to gridding limitations in Delft3D, three channels were not completely discretized (A-C). The inlet boundary condition is specified at the upstream extent of the channel west of bifurcation 1.1 and the 18 outlet boundary conditions are specified at the termination of each channel.
4.3 Numerical modeling approach

The numerical model is designed to imitate channel processes on the Mossy delta in Saskatchewan, Canada (Figure 4-1). The delta started forming 75 years ago after the sediment wedge from the 1870s avulsion of the Saskatchewan River reached the shores of Lake Cumberland [Smith, et al., 1998]. Currently, the delta is composed of a network of channel bifurcations not directly impacted by human activity. Indirectly, the delta is affected by a hydroelectric dam 100 kilometers upstream, which went into operation in 1963. Flood discharges have been reduced and daily releases cause fluctuations in water level at the delta head of a few centimeters. Analysis of delta growth using serial aerial photos and 94 boreholes taken in 1995 indicates that pre-dam sediment delivery to the delta starting in 1947 amounted to approximately 6 x 10^9 kg yr^{-1} [Oosterlaan and Meyers, 1995]. Post-dam delivery has been reduced to approximately 2 x 10^9 kg yr^{-1}, a threefold decrease. For the experiments we used field data from the Mossy delta collected during the summers of 2006-2007 as boundary and initial conditions. The channel network was subjected to perturbations by closing bifurcate channels of varying discharge and location within the network, and the response was documented.

We conducted two sets of experiments. The first set is aimed at determining if the Mossy delta channel network is in equilibrium at present, as has been suggested by other workers. The second set is aimed at determining how the equilibrium delta network of the Mossy delta responds to perturbations. We subjected the equilibrium network to perturbations by closing different channels through a range of bifurcation orders and discharges. For each experiment where a channel was closed we catalogued
the effect by looking at the how the discharge, water surface, and bed elevations throughout the network changed. In particular we focused on which, if any, channels in the network closed, which channels took the increased flow, and how the bifurcations responded and came to a new equilibrium. In all experiments channels were allowed to deepen, shoal, and narrow in response to a perturbation, but we did not allow avulsions, channel widening, overbank flow, or delta progradation.

4.3.1 Numerical model description

Delft3D v. 3.27 is used to model the morphodynamic evolution of the distributary channel network. Delft3D simulates fluid flow, sediment transport, and morphological changes at time scales from seconds to years and has been validated for a wide range of hydrodynamic, sediment transport, and scour and deposition applications in rivers, estuaries, and tidal basins [Hibma, et al., 2004; Lesser, et al., 2004; Marciano, et al., 2005; van Maren, 2005]. The equations of fluid and sediment transport and deposition are discretized on a curvilinear finite-difference grid and solved by an alternating-direction implicit scheme. An advantage of Delft3D is that the hydrodynamic and morphodynamic modules are fully coupled; the flow field adjusts in real time as the bed topography changes.
4.3.2 Governing equations

Delft3D solves the non-uniform, unsteady, incompressible Reynolds-averaged Navier-Stokes equations for fluid flow under the shallow-water and Boussinesq assumptions. The equations consist of conservation of momentum, conservation of mass, and the transport equation. All experiments are 2D, depth integrated, and the horizontal eddy viscosities are defined using a horizontal large-eddy simulation that relates the horizontal fluid shear stress to the horizontal flow velocities. The horizontal large-eddy simulation uses a quasi-2D sub-grid scale turbulence model. The sub-grid scale turbulent properties, such as the eddy viscosity, are calculated by passing the velocity field through a high pass filter to remove the large eddies from entering the sub-grid scale model. Delft3D recommends that the high pass filter be set to remove eddies with a period that scales with the largest eddy; for our simulations we used an eddy period of 10 seconds, which is comparable to measured eddy periods that span the flow depth in sand-bed rivers \[Best, 2005\]. Additional tests proved that the model results were generally insensitive to our choice of eddy period.

All results presented here use the vertically integrated 2D equation set in Delft3D because the 2D simulation accurately predicts the discharge in reaches throughout the channel network. This indicates that the processes responsible for distributing water and sediment at a bifurcation can be captured in two dimensions. While certain three-dimensional processes, such as secondary circulation and vertical concentration gradients of suspended sediment, may play a role in sediment and water distribution at a bifurcation, a 3D solution with five equally-sized computational layers in the vertical did not significantly differ from the discharge distribution in a 2D
solution (Figure 4-2). This suggests that the additional processes captured in a 3D model are of secondary importance to distributing water and sediment at a bifurcation.

Delft3D has separate mathematical treatments for the erosion and deposition of cohesive and noncohesive sediment.

Cohesive sediment is defined as silt-sized and finer, whereas noncohesive sediment is defined as sand-sized and coarser. The formulation for cohesive sediment erosion and deposition is based on the formulation of Partheniades [1965] and Krone [1962], whereas the formulation for noncohesive sediment erosion and deposition is based on the Shields curve.

Cohesive and noncohesive sediment can be transported as bedload or suspended load depending on the grain size and the flow strength. Bedload transport rate per unit width is calculated from van Rijn [1984]. The magnitude and direction of the bedload transport vector is adjusted for favorable and adverse longitudinal slopes according to Bagnold [1966] and for transverse slopes according to Ikeda [1982]. Suspended load transport rate is calculated by solving the vertically-integrated three-dimensional diffusion-advection equation, where the sediment eddy diffusivities are a function of the fluid eddy diffusivities. The fluid eddy diffusivities are calculated in the equations for

Figure 4-2 Calculated versus observed discharge (Q) in various reaches of the Mossy delta. The match between the observed and calculated does not change for a 3D simulation.
fluid flow using the horizontal large eddy simulation and grain settling velocity. The horizontal large eddy simulation effectively captures the spatial variation in horizontal eddy diffusivity known to exist in rivers. Gradients in the sediment transport vectors are used to determine changes in bed topography using the Exner equation. For a more detailed discussion on the mathematics of Delft3D and the flow/topography interactions see Lesser et al. [2004].

4.4 Numerical modeling setup

4.4.1 Grid considerations

A numerical grid of the active distributary channel network (Figure 4-1) was constructed using RGFGRID in Delft3D v. 3.27. The planform configuration of the grid was based on a composite aerial photograph of the Mossy delta from 2003. The composite photograph was constructed from a series of forty oblique aerial photographs flown by the Information Services Corporation of Saskatchewan [2003]. The oblique images were georectified using ArcGis v. 9.1 by noting static features like bedrock outcrops, lake shorelines, and tree locations, and matching them to a professionally georectified 1973 map of the delta produced by the Canadian government. The forty 2003 georectified images were then stitched together in ArcGis v. 9.1, producing a composite georectified 2003 photo which is accurate to approximately +/- 5 meters. Accuracy was determined by comparing the final 2003 georectified image to the 1973 image.
The grid was constructed by tracing the channel margins from the 2003 composite photograph in the Delft3D gridding software. The positions of channel margins in 2003 were cross-referenced with 2006 aerial photographs taken during a reconnaissance of the delta. This was done to insure that the locations of the channel margins were internally consistent with the initial- and boundary-conditions data that were collected in 2006 and 2007. In one case, the 2003 photographs did not cover the extent of the delta (see south of bifurcation 3.4 in Figure 4-1) and individual photographs from 2006 were georectified so that the channel extent could be confidently located.

Distributary channels were discretized from the head of the delta to the shoreline. To maintain computational efficiency, only the channelized portions of the network were included in the grid and therefore floodplain processes are not be considered in this study. Discretization was terminated where the channel becomes subaerially unconfined in the lake as determined from the 2003 aerial photograph. Minor channels less than 10 meters wide were not included in the grid because they would be approximately two grid cells wide, and moreover they were too narrow and shallow to collect boundary and initial condition data. Also, two channels near bifurcation 1.1 at the head of the delta were not discretized because we did not have any initial or boundary condition data (Figure 4-1) and channels A-C in Figure 4-1 were eliminated due to limitations of the Delft3D gridding software. The error introduced by ignoring these channels is discussed below.

To insure that numerical results are independent of grid cell size [Hardy, et al., 2003] we conducted preliminary experiments with double the number of cells in the
grid. Results from the two resolutions produced the same flow field, leading us to conclude that our results are insensitive to the choice of grid cell size. Therefore, we chose a numerically efficient grid with approximately 15,000 cells that still resolves topographic details in the evolving system. The resolution of the grid varies across-channel and down-channel and is generally denser around bifurcations to insure an accurate division of water and sediment (Figure 4-1). Grid cells range in shape from rectangles to squares and the cell size ranges from 4 by 4 meters to 6 by 16 meters. To determine the appropriate time step, a series of sensitivity tests was conducted where the time step was reduced until the model converged on an answer. The time step used in all experiments for this study is 6 seconds.

Grids in Delft3D should be smooth and each cell should be orthogonal in order to conserve mass and momentum. To achieve orthogonality through curved channel reaches and around the bifurcation points, extra grid cells were added and an orthogonal transformation was applied using the Delft3D gridding software. The computational grid used in these experiments has a maximum deviation from orthogonality of approximately 20 degrees. This orthogonality does not affect the solution; grids with lower orthogonality achieved results similar to results computed with the grid in this study.
4.4.2 Boundary and initial conditions

4.4.2.1 Field data collection

Data for boundary and initial conditions were collected during the summers of 2006 and 2007, including river bed topography, hydraulic data, water surface elevations, and grain size data over the entire delta at near bankfull stage. Bed topography was collected using an EAGLE FishElite 500c single-beam echo sounder. Discharges at 25 cross sections distributed throughout the net were measured with a Teledyne RD instruments Rio Grande acoustic Doppler current profiler (ADCP). Typical ADCP discharge measurements for systems similar to the Mossy delta are accurate to within 1 to 7% of the true discharge [Oberg, et al., 2007] and that accuracy depends upon the duration of the cross-section measurement. Our discharge measurements have an accuracy to within 10%, determined by comparing the discharge in an upstream reach to the cumulative discharge in the bifurcated downstream reaches. Our error is higher than published values because the some channels are shallow (~1 meter) and therefore the fixed blanking depth of the ADCP (5 to 10 centimeters) over which no data are collected became a larger percent of the flow depth in shallower channels. The ADCP was located spatially and temporally using a Leica differential global positioning system (DGPS) in real time kinematic mode, which produced accuracy in relative position (DGPS base station to mobile rover) of a few millimeters in the horizontal and vertical positions. The boat velocity and track position along the survey lines were monitored and were held as constant as possible during surveying. For each cross-section two repeat surveys were made and the results were averaged.
The water surface slopes through the channel network were measured by mounting a survey grade DGPS rover (receiver) in a boat and floating down the middle of the channels recording water surface elevations every second. The water surface elevation data were processed with Ski-Pro v. 3.0 using a base station of known elevation. This technique is advantageous because the high temporal resolution and vertical accuracy resolve reach scale changes in the water surface. To validate this technique, we floated the same river reach multiple times over different days and observed the same features of water surface topography in each float.

Bed grain size data were sampled with a standard bed grab sampler. Suspended sediment concentrations were measured with a standard US DH-48 depth integrated suspended-sediment sampler.

4.4.2.2 Values of boundary and initial conditions

The numerical grid has one inlet and 18 outlets. The inlet boundary condition at the head of the delta is a steady, uniform discharge of 250 m³/s. We use 250 m³/s, instead of the 295 m³/s measured at the head of the delta because: 1) there is 35 m³/s of flow determined by ADCP measurements leaving the main channel through two minor channels on the northern bank up and downstream of bifurcation 1.1 (Figure 4-1); and 2) the discharges as measured by the ADCP in the northern and southern branches of bifurcation 1.1 sum to 250 m³/s. Sediment flux at the inlet is in equilibrium with the flow and consists of two grain sizes: a cohesive fraction of silt and a noncohesive fraction of fine-grained sand. Initially in the model domain there is 2.5 meters of evenly mixed erodable substrate, of which 0.5 meters is noncohesive and 2 meters is
cohesive. These values were chosen to reflect the average substrate thickness and composition in the delta [Oosterlaan and Meyers, 1995]. Equilibrium fluxes of these two sizes at 250 m³/s water discharge results in a yearly delivery of 2x10⁹ kg of sediment to the delta consistent with the observed growth from aerial photos.

The outlet boundary conditions are steady, uniform water surface elevations of 266 meters, the average elevation of Lake Cumberland. In the case of the two channels that are not discretized to the delta shoreline (A,B Figure 4-1), the steady boundary condition was a water surface elevation measured in the field. Even though the water surface elevations are fixed, the bed elevations at the downstream boundaries are allowed to adjust during the simulations.

The initial bed topography used in the modeling experiments is from bathymetric data collected over the entire delta. For bifurcations 1.1, 2.1, 2.2, 3.1, 3.3, 3.4, 4.2, and 5.1 the bed topography was mapped in detail to resolve all the topographic features. For the remaining channel reaches bed topography was measured at cross-sections spaced every 150 meters and the cross-sections were linked together by measuring the bed topography through the channel centerline. The elevations of sand bars above the water surface were estimated in the field. The bed topography data were imported into Delft3D and triangularly interpolated to the computational grid.

Roughness values were calculated using the Chezy roughness formulation. We used field observations of bed state to calculate Chezy coefficients for each channel segment. The Chezy coefficients used in the model range from 20 to 90 m¹/²/s, consistent with our observation of large dunes in some reaches and plane beds in others.
Delft3D allows the user to speed up the bed adjustments by multiplying the deposition or erosion rate in each time step by a morphological scale factor. A series of sensitivity experiments showed that the final solution is insensitive to a morphological scale factor less than 1000. We used a factor of 500.

4.5 Results and discussion

4.5.1 Equilibrium configuration of the Mossy delta channel network

Edmonds and Slingerland [2007; 2008] argued that the Mossy delta distributary network presently is in equilibrium with its incoming discharge and sediment supply because the channels conform to the hydraulic geometry scaling laws [Edmonds and Slingerland, 2007] and of the eight bifurcations tested, all conform to the stability criteria for fine-grained bifurcations [Edmonds and Slingerland, 2008]. As an additional test we set up a fully morphodynamic numerical experiment that uses the field data as initial and boundary conditions to replicate the conditions on the Mossy delta during the summers of 2006 and 2007. The initial and boundary conditions must be close to the equilibrium solution because if the initial guess is not accurate the nonlinear nature of the equations causes the system to runaway to an unstable solution. Therefore to find the equilibrium solution with the measured initial and boundary conditions we slightly adjusted (i.e. +/- 5% of the measured value) the roughness and bed topography in reaches where data were sparse or not well constrained (e.g. the tips of the distributary channels) searching for a Delft3D solution that minimizes the
deviations between predicted and observed discharges and water surface elevations in the various reaches.

Results show that there is an initial flux of sediment out of the delta network as the model reworks the bed topography (Figure 4-3). After 17 years of computation time an equilibrium solution is found wherein the network has adjusted its water surface slope and bed topography such that the water and sediment entering at the head is passed through and the average value of the absolute value of the changes in bed elevation (|ΔE|) over one year approaches zero (Figure 4-3). When equilibrium is reached most of the original bed topography has been reworked and there is still approximately 0.5 to 1.5 meters of erodable sediment at each cell in the model domain. In natural systems, equilibrium deltaic systems are net depositional because the downstream boundary is changing as the delta progrades or sea level rises; however, because we have simplified our experiments and ignored the effect of delta progradation we argue that fluvial grade is an appropriate measure of equilibrium for a static

![Figure 4-3](image)

**Figure 4-3** After approximately 17 years of computation time the average of the absolute values of the differences in bed elevations between time steps goes to zero and remains there for many multiples of non-dimensional time ($T_{ND}$) indicating that the Delft3D solution is in equilibrium. See text for definition of $T_{ND}$. 

88
network.

We consider this to be an equilibrium solution in the sense that the network remains at grade for many multiples of the morphological time scale ($T_m$) [Miori, et al., 2006], where

$$T_m = \frac{W_a D_a}{q_{sa}}$$  \hspace{1cm} (4)

$W_a$ is width (m), $D_a$ is the depth (m), and $q_{sa}$ is the sediment transport rate per unit width ($m^2/s$) of reach $a$. $T_m$ is the duration over which the amount of sediment needed to fill a cross-section is transported through that cross-section. It is a dynamic equilibrium solution if there is suspended and bedload transport in all reaches and the change in the summation of $|\Delta E|$ with time varies by no more than 1% around zero for at least 15 multiples of nondimensional time ($T_{ND}$), where

$$T_{ND} = \frac{T}{T_m}$$  \hspace{1cm} (6)

$T$ is the total time elapsed (days), and $T_{ND}$ is the nondimensional time, or the multiples of the morphological time scale elapsed during the computation. The solution reaches equilibrium after 17 years and remains there for 3 years (approximately 60 multiples of the average $T_m$) until the simulation is stopped. To estimate $T_m$ for the Mossy delta we used spatially averaged values of sediment flux and the channel cross-section. The average $T_m$ for the Mossy delta is approximately 22 days.
4.5.1.1 Comparison of the equilibrium solution to field data

The computed equilibrium solution for the Mossy delta compares favorably to field data. The computed discharge in each bifurcate channel is similar to the observed discharges in the delta (Figure 4-2) with an $R^2 = 0.88$ and a slope close to 0.9. This is strong evidence that the model is simulating the natural processes at bifurcations responsible for dividing water and sediment asymmetrically. Furthermore, the intricacies of the water surface slopes that are produced as water flows through a bifurcation are captured by the model (Figure 4-4 and Appendix A, Figure A-12).

The equilibrium solution also makes some interesting predictions about the Mossy delta. The model predicts that the south channel in bifurcation 4.4 and the south channel in bifurcation 3.3 will be abandoned (Figure 4-5). These predictions are consistent with the observations of abundant vegetation throughout each channel, which suggests low flow velocities and lack of sediment transport.

Figure 4-4 Water surface elevations calculated by the model compare well with field data. Notice that the model correctly predicts a steeper water surface slope downstream of the bifurcation point. For location see dotted line in Figure 4-5.
Also, the equilibrium solution predicts that there is more cumulative sediment transported through the northern half of the delta (0.0166 m³/s of sediment) than through the southern half (0.007 m³/s), where the northern half is defined by all the channels downstream of the north branch of bifurcation 1.1. This suggests that the long term evolution of the delta will be dominated by channel progradation in the northern

Figure 4-5  Predicted equilibrium water depths in distributaries of the Mossy delta after 17 years of computation. The area of the brown circles at the channel mouths is proportional to the predicted sediment flux exiting at that location. Black represents parts of the delta channels that are subaerial. Two channels (A and B) have been nearly abandoned; only a small discharge is flowing through them and there is no sediment transport.
half. This is curious because the delta has maintained approximate symmetry in plan-view during progradation since its initiation in 1930. Deltaic deposits are thicker in the north [Oosterlaan and Meyers, 1995], implying that accommodation space is greater there. It is possible that the distributary net has developed to maintain a radial symmetry, compensating for the variable accommodation space by varying its sediment flux to the coast. Alternatively, this variation could be part of an autocyclic adjustment in which one side grows forward and then negative feedbacks shift the locus of sedimentation to the other side. One possible feedback is that the path of maximum sediment discharge changes through time and corresponds to the shortest path from the head of the delta to the shoreline and the location changes as channels lengthen at different rates. While this is reasonable, it does not explain the current path of sediment flux on the Mossy delta. The shortest path from the head of the delta to the shoreline travels through the south branch of bifurcation 2.2 and the middle branch of 3.4 (total distance of ~4100 m), yet field data and the model predictions indicate that this is not the location of maximum water and sediment discharge. The path of maximum sediment and water discharge from the delta head travels through the north branch of bifurcation 2.1, the south branch of 3.1 and the north branch of 3.2, and is 1100 meters longer.

In summary, we take Figures 4-2 through 4-4 as evidence that the equilibrium form of the Mossy delta is computable in Delft3D and therefore we can test the stability of this form by perturbing the channel network and comparing the perturbed solution to the equilibrium solution.
4.5.2 Perturbing the equilibrium configuration of the Mossy delta

To understand how a delta channel network responds to perturbations, we perturbed the equilibrium channel network of the Mossy delta by closing various bifurcate channels (Table 4-1). The experiments are abbreviated such that B43N refers to the closure of the north branch of bifurcation 4.3. The channels were chosen because they span a variety of discharges and locations on the delta.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>time to equilibrium (years)</th>
<th>water discharge of closed channel (m³/s)</th>
<th>sediment discharge of closed channel (m³/s)</th>
<th>Cross-sectional area of closed channel (m²)</th>
<th>north (N) or south (S)</th>
</tr>
</thead>
<tbody>
<tr>
<td>B22S</td>
<td>17</td>
<td>21</td>
<td>0.00198</td>
<td>47.2</td>
<td>S</td>
</tr>
<tr>
<td>B31S</td>
<td>17.8</td>
<td>64</td>
<td>0.00602</td>
<td>87.1</td>
<td>N</td>
</tr>
<tr>
<td>B34M</td>
<td>7.7</td>
<td>9</td>
<td>0.0009</td>
<td>12.3</td>
<td>S</td>
</tr>
<tr>
<td>B32N</td>
<td>6.8</td>
<td>4</td>
<td>0.000385</td>
<td>8.7</td>
<td>N</td>
</tr>
<tr>
<td>B42N</td>
<td>18.2</td>
<td>52</td>
<td>0.00488</td>
<td>58.5</td>
<td>N</td>
</tr>
<tr>
<td>B42S</td>
<td>9.3</td>
<td>13</td>
<td>0.00114</td>
<td>34</td>
<td>N</td>
</tr>
<tr>
<td>B43N</td>
<td>10.4</td>
<td>50</td>
<td>0.005</td>
<td>41.7</td>
<td>N</td>
</tr>
<tr>
<td>B51N</td>
<td>6.5</td>
<td>12</td>
<td>0.000986</td>
<td>27.5</td>
<td>N</td>
</tr>
<tr>
<td>B61S</td>
<td>7.3</td>
<td>5.5</td>
<td>0.00042</td>
<td>8.1</td>
<td>S</td>
</tr>
</tbody>
</table>

Table 4-1 The nine perturbation experiments conducted in this study. The last column refers to whether the closed channel is in the northern or southern half of the delta.

The initial and boundary conditions for the nine perturbation experiments were taken from the previously computed equilibrium configuration. To insure that the results of the perturbation experiments are not due to improperly restarting the simulation, we conducted some control experiments in which runs with no changes to the domain were restarted from the equilibrium configuration. The control experiments exactly reproduced the equilibrium configuration. To simulate closure of a channel an infinitely thin barrier was placed between the two adjacent cells at the entrances to
various channels on the delta. When the perturbation experiments are restarted the channel is instantly closed due to the presence of the barrier. In all experiments the thin dam is placed orthogonally across the entrance of the channel. We did experiments with the dam oriented obliquely to the entrance of the channel. In both cases the response of the network was the same, indicating that the results are not dependent on the orientation of the thin dam. This instantaneous closure most closely simulates an ice or log jam.

Similar to the equilibrium experiment, each perturbation experiment was computed until the average $|\Delta E|$ over one year approaches zero. Running an experiment until the average $|\Delta E|$ is actually zero can take considerable computational time because the approach to zero is asymptotic. To save computational time, the perturbed experiments are considered to be in equilibrium when the average $|\Delta E|$ over 20 to 30 years of computational time because the approach to zero is asymptotic. To save computational time, the perturbed experiments are considered to be in equilibrium when the average $|\Delta E|$ over

![Graph showing average $|\Delta E|$ over time after perturbation.](image)

Figure 4-6  All nine perturbation experiments come to a new equilibrium after a channel is closed. The larger the discharge of the closed channel the larger the initial $|\Delta E|$ in the experiments. In general, the experiments reach equilibrium after 20 to 30 years of computational time. The experiment names are abbreviated so that B22S refers to the closure of the south branch of bifurcation 2.2. See table 4-1 for the precise times to equilibrium.
one year is less than 0.0025 meters (Figure 4-6). We chose this number because it is below the detectable limit of bed level changes in the field.

The modeling experiments presented here are simplified and do not allow channels to increase their top width. Therefore, the responses of the network to channel closure apply to delta networks at time periods less than the channel width response time. Serial aerial photographs from the Mossy delta suggest that channel width adapts to changes in discharge in approximately 10 to 20 years. Therefore, the results here strictly apply to time periods less than 10 to 20 years.

Results show that closing a channel in a delta channel network causes: 1) reorganization of the distribution of water and sediment discharge throughout the delta channels; 2) changes in the number of active channels within the delta network; and 3) possible long term changes in the evolution of the delta.

4.5.2.1 Reorganization of water and sediment discharge

After the network is perturbed it comes to a different equilibrium that is characterized by a new distribution of the water and sediment discharge throughout its branches (Figure 4-7). Thus, closing a channel of any size or location on the delta reorganizes the discharge over an area larger than the closed channel and its associated bifurcate channel. This reorganization is an interesting response because only a proportion of the discharge from the closed channel flows down the other bifurcate branch (e.g. experiment B51N in Figure 4-7); the remaining discharge is distributed among channels upstream of the closed channel. This occurs because the water upstream of the closed channel rises to generate enough head to drive an increased flow
through the (now) undersized remaining bifurcate channel (A in Figure 4-8). If a
significant increase in depth is needed, the effect propagates far upstream, in some cases
as far as the head of the delta. If there are additional bifurcation points upstream of the
closed channel, then as the water depth increases to generate more head, the water
surface elevation increases at the bifurcation point causing additional discharge to flow
down that channel (B in Figure 4-8). Therefore, the discharge from the closed channel
is distributed over more channels upstream (Figure 4-7). In most experiments the
highest percent change in water and sediment discharge occurs in the first bifurcation
upstream of the one closed because this is where the rise in water surface elevation is
the highest.

The mechanics of this response can be used to predict how certain reaches of the
Mossy delta network respond to channel closure. In most reaches of the delta the
discharge increases (Figure 4-7). However, in the connected channel reaches that are
each progressively farther upstream of the closed channel (e.g. bifurcation reaches 4.2
to 3.1 to 2.1 to 1.1 in Figure 4-7 B42S) there is always a decrease in water and sediment
discharge. This occurs because the water surface elevation in those reaches rises to
generate the additional head needed to pass discharge through the undersized bifurcate
channel and therefore discharge is lost from those reaches as it is routed down other
branches at bifurcation points. In all other reaches, the water discharge increases
because the water surface elevation at the bifurcation point increases.
Figure 4-7  Closing one channel causes reorganization of the water discharge throughout the delta network. The colored squares represent the percent changes in discharge through a channel reach after a channel is closed. We consider changes less than 1.5% to be zero because they would not be detected by an ADCP. The dashed line in each image is the channel reach that was closed. The experiment names are abbreviated so that B22S refers to the closure of the south branch of bifurcation 2.2.
Figure 4-8 Water surface elevation of bifurcations 3.1, 4.1, and 4.2 (see Figure 4-1 for location within the delta). Closing the north channel of bifurcation 4.2 causes an increase in head as additional discharge flows into the southern channel. This increase in head migrates upstream after the channel is closed (A). The increase in head travels past bifurcation 3.1 and increases the water surface elevation at the bifurcation point, which increases the discharge down that channel (B). This is the mechanism that distributes the effect of the channel closure over most of the delta network.
The extent of the reorganization of water and sediment discharge after closure scales with the discharge of the closed channel and the proximity of bifurcation points to the perturbation. When large discharge channels are closed (e.g., B31S, B42N, and B43N in Figure 4-7) more reaches of the delta are affected than if smaller discharge channels are closed (e.g., B34M and B61S in Figure 4-7). But location of the closure matters too, as for example, in experiments B42S and B51N where the closed off channels have approximately the same discharge. The extent of the reorganization of the network is significantly different (Figure 4-7) even though the distance upstream that the water surface elevation is affected is approximately 3.5 kilometers along the channel centerline for each case. From the closed channel B42S that distance reaches bifurcation 1.1, whereas from B51N that distance is short of bifurcation 2.1. Therefore the extent of the perturbation in experiment B42S is much larger because of the proximity of other bifurcations.

4.5.2.2 Changes in the number of active channels

In some cases closing a channel changes the distribution of water and sediment discharge enough that a previously active bifurcation becomes abandoned or reactivated. In experiments B22S, B31S, and B42N closure causes the north branch of bifurcation 5.1 to decrease in discharge by over 50% and eventually causes the channel to be abandoned (Figure 4-7). In experiment B61S closing the channel reactivates the previously abandoned bifurcation 4.4 by increasing the water surface elevation at its
bifurcation point. Yet relative to the cumulative channel length of the delta, these are small changes and we conclude somewhat surprisingly, none of the perturbation experiments significantly reduces the number of active channels in the network.

We think this delta network remains stable after a closure because the changes in Shields stress at the bifurcations are small. Theoretical work has shown that the equilibrium discharge ratio of two bifurcate channels is a function of the Shields stress in the unbifurcated upstream reach [Bolla Pittaluga, et al., 2003; Bertoldi and Tubino, 2007; Edmonds and Slingerland, 2008]. Edmonds and Slingerland [2008] looked at the stability of fine-grained cohesive bifurcations like those of the Mossy delta, and found for a given channel roughness and aspect ratio, that cohesive bifurcations are in equilibrium over a range of Shields stresses from 1 to 2.5. The average Shields stress in experiment B43N is 1.2 N/m² and the percent change in Shields stress after the closure of the channel ranges from -20 to 38% over the delta. Therefore the variation in Shields stress, according to the theoretical results of Edmonds and Slingerland [2008], is not enough to cause a bifurcation to become unstable. It is possible that a large enough perturbation could significantly change the number of active channels in a delta, but currently there is no work that defines what size or type of perturbation is necessary.

The direction of the change in Shields stress can be used to predict the response of the bifurcation. Edmonds and Slingerland [2008] found that the equilibrium discharge ratio of the bifurcate channels increases with increasing Shields stress of the upstream reach. That study isolated the bifurcation from the network and focused on bifurcations with equal water surface elevations at the downstream boundaries of the bifurcate channels. The same general response is seen for a network of bifurcation
(Figure 4-9). For eight bifurcations on the delta, an increase (decrease) of the Shields stress produced an increase (decrease) of the equilibrium discharge ratio (Figure 4-9).

4.5.2.3 Implications for the long term evolution of a delta channel network

Even though the effect of closing one channel does not significantly change the number of active channels, it does reorganize the distribution of water and sediment discharge throughout the network significantly enough to alter the long term evolution of the delta. At equilibrium, 70% of the sediment discharge presently exits through the channels in the northern half of the delta. After the closure of a channel, no matter where it is located, the balance of sediment flux between the north and south half changes (Figure 4-10). The change is not enough to reverse or equalize the sediment
fluxes of each half, but it may be enough to change the long-term evolution. For example, in experiment B42N the sediment flux exiting from the south half of the delta increased by 15% (Figure 4-10). Over a period of one year this results in an additional $3 \times 10^4$ m$^3$ of sediment deposited basinward of the delta front, thereby leading to faster channel progradation and river mouth bar construction. In the southern half of the delta ~90% of the sediment is transported through the north channel of bifurcation 5.2. Assuming a river mouth bar has an average volume of 8500 m$^3$ for that bifurcation order and that all the sediment transported as bedload is deposited as a river mouth bar, the bar will be subaerial 10 years sooner because of the change induced by the perturbation. This has the potential to change the dynamics of the system, especially if the presence of the bar causes a morphodynamic backwater wave that could lead to avulsion as suggested by Hoyal and Sheets [2008] and Edmonds and Slingerland (in review).

The direction of the change in sediment flux between the north and south branches depends on where the perturbation is located. If the abandoned channel is in the north half of the delta than the water surface elevation increase originates from that half to

![Figure 4-10](image)

**Figure 4-10** The percentage of sediment flux entering the head of the delta that is discharged out of the north and south halves at equilibrium (Eq) and the other perturbation experiments. The (n) or (s) signifies whether the closed channel is in the north or south half of the delta.
generate enough head for the water to flow through the undersized bifurcate channel. This results in a water surface elevation rise at bifurcation 1.1, which leads to more water and sediment transported down the south branch of bifurcation 1.1 and therefore the south half of the delta (Figure 4-10). The opposite is true if the channel is closed in the southern half of the delta.

4.5.3 How will real delta networks respond to perturbations?

The results presented here provide insight into how one particular delta channel network responds to closure of a channel. It represents a purely distributive end-member where few channels reconnect. In other delta systems, such as the Lena River delta, there are many interconnected channels that could change the response of the network (Figure 4-11). If a bifurcate channel were abandoned on the Lena delta, the effect would likely be distributed over a smaller area because the high number of

![Figure 4-11 Aerial image of the Lena River delta from Landsat 7 satellite. Closing a channel in the Lena delta would spread the discharge over a smaller area (shown as blue arrows) because the interconnected channel network would dissipate the perturbation over a smaller area.](image-url)
interconnected channels would dissipate the change in water surface elevation before it traveled up the delta.

These results also provide insight into how a delta channel network evolves as it progrades basinward. Results from this study show that if a channel is abandoned on one half of the delta, more sediment will be routed to the other half (as is the case in Figure 4-10), which may influence the long term evolution. For example, currently on the Mossy delta the majority of sediment is being delivered to the northern half. But, in 1977 nine of the fifteen newly created bifurcations at the delta shoreline were in the southern half suggesting that more sediment was being delivered there. By 1982, five of those nine bifurcations in the southern half of the delta (compared to two in the northern half) were abandoned. The abandonment of more bifurcations in the southern half of the delta may have caused more sediment to be delivered to the north half. This may be an important natural feedback in delta systems that maintains symmetrical delta progradation through time. If one half of the delta is prograding faster and constructing more bifurcations per unit time, then statistically more bifurcations on that half of the delta would be abandoned, which would cause additional sediment to be routed to the other half of the delta and equalize the sediment distribution over the shoreline.

4. 6 Conclusions

We have attempted to determine the far-field effects of various channel closures on a delta channel network through numerical modeling experiments simulating the
Mossy delta of the Saskatchewan River. In general the distributary network was relatively stable. Although significant amounts of water and sediment were re-routed, most channels and bifurcations adjusted bed elevations to remain open. In a few experiments channels at the shoreline were abandoned or reactivated but these were of low discharge and high order.

This stability is all the more surprising because the effect of a closure is felt over the entire delta. In all nine perturbation experiments the water and sediment fluxes from the closed channel are redistributed among other channels throughout the delta. This occurs because as the additional water flows into the other, undersized bifurcate channel, the water depth upstream increases to generate enough force. The increase in depth upstream can propagate as far as the head of the delta and if there are other bifurcation points upstream of the closed channel then more discharge will routed down those channels on the delta.

The extent of the effect of closing a bifurcate channel depends upon the discharge of the closed channel and the proximity of other bifurcations. Closing higher discharge channels will cause more discharge to flow into the other bifurcate channel, which will require a larger increase in water depth upstream that will propagate farther and therefore affect more parts of the delta.

Closing a channel can also affect the long term evolution of the delta. Closing a channel on the south half of the delta increases the sediment flux to the north half and vice versa. In the most extreme cases, this can cause a 5-7% swing in the sediment flux from one half of the delta to other. Simple calculations show that this swing in sediment flux can speed up the formation of a river mouth bar by ten years.
Future work should include understanding what effects variable channel width, floodplain processes, and delta progradation have on perturbations to the delta channel network.
Chapter 5: Conclusions
5.1 Summary and synthesis

Deltaic landforms, especially those dominated by fluvial processes, tend to exhibit a common morphology and branching pattern. Therefore, if a link between the dominant process and morphology can be established, then river-dominated delta morphology can be predicted for a variety of different boundary and initial conditions. Developing an ability to predict the behavior and morphology of deltas has important implications for protecting human population, wildlife, and infrastructure. Earlier work had shown how river-dominated deltas are built by bifurcations around river mouth bars. However, that work had not addressed the three important questions that I asked in Chapter 1: 1) What processes, other than river mouth bar deposition, participate in the formation of river-dominated delta channel networks; 2) Once formed, what are the equilibrium states of the delta channel network; and 3) How stable are those states to perturbations? These questions were addressed in Chapter 2-4, respectively.

In Chapter 2 I looked at processes that participate in the formation of delta channel networks. Using physical models of deltaic systems, this chapter documented that the process of avulsion in deltas may be important in network growth. Previous work on delta network growth, which has emphasized the role of bifurcation around river mouth bars, had not considered the contribution of the avulsion process. I showed that avulsions in deltas are caused by a growing river mouth bar at the shoreline. The stagnated mouth bar triggers a wave of bed aggradation moving upstream that increases cross-levee flows and bed shear stress. An avulsion occurs as a time-dependent failure of the levee where the largest average bed shear stress has been applied for the longest time. Most notable is that I was able to predict the avulsion location with a high degree
of certainty suggesting that conceptualization of avulsion as a time-dependent failure of
the levee, rather than a threshold, will aid in wetland restoration on deltas and may
improve avulsion prediction in other fluvial systems. Furthermore this study
demonstrates that a delta channel network is constructed by avulsions in addition to
channel bifurcation via river mouth deposition, and therefore to accurately predict delta
evolution future models must include both processes of bifurcation.

In Chapter 3 I explored the equilibrium states of bifurcations once they have
formed. Using Delft3D, I conducted numerical experiments that defined the
equilibrium configurations of fine-grained, cohesive bifurcations. The results show that
over a range of channel aspect ratios, friction factors, and Shields numbers, there are
three equilibrium functions relating the discharge ratio of the bifurcate arms at
equilibrium to Shields number. One function defines symmetrical configurations (equal
partitioning of discharge), while the other two define asymmetrical configurations
(unequal partitioning of discharge). When subjected to a perturbation, such as a small
sediment mound in a bifurcate channel, the asymmetrical bifurcations return to their
equilibrium configuration whereas the symmetrical bifurcation moves to an
asymmetrical stable equilibrium solution. These results suggest that asymmetrical
bifurcations are prevalent in nature because they are stable to a wider range of
perturbations. The model results agree well with field data collected on the bifurcations
of the Mossy delta; the field bifurcations contain remarkably similar asymmetric bed
geometries and water surface profiles to those predicted by Delft3D.

Chapter 4 examined how stable a delta channel network is to perturbations. Even though chapter 3 showed that single bifurcations are stable to perturbations, it is
not necessarily true that a network of bifurcations is stable to perturbations because they are connected via bifurcation nodes that transmit information throughout the delta network. I conducted a suite of numerical modeling experiments to test this idea. I conducted nine experiments that were designed to replicate the Mossy delta. In each experiment, the delta network was perturbed by closing a bifurcate channel. Results show that when a bifurcate channel is closed the sediment and discharge fluxes from that channel are redistributed throughout the entire delta. The extent of that effect is a function of the discharge of the closed channel and the proximity of other bifurcations.

The preceding chapters demonstrate two general characteristics about river-dominated deltas. First, these deltas are a dynamic environment with positive and negative feedbacks that act to keep the delta near equilibrium. For example, if one channel captures most of the flow and sediment it will construct a mouth bar offshore that will generate a wave of sedimentation that moves upstream through the channel network. That wave of sedimentation causes overbank flow, which triggers an avulsion thereby redistributing the flow and sediment across the delta. Second, river-dominated deltas develop such that bifurcations route sediment and water asymmetrically in their downstream branches. This is preferred to a symmetrical division because it is stable to perturbations. At a broader scale, river-dominated delta channel networks are stable to perturbations, such as the closing of a channel. But, surprisingly depending upon the discharge of that channel and the proximity of other bifurcations the effect of its closure can significantly alter the distribution of water and sediment throughout of the delta to the point that long term evolution may be changed.
These results have broad implications for fluvial geomorphology and other fields of geology as well. Firstly, the results presented here all move toward developing a predictive model of delta behavior. Therefore, hazards associated with delta environments, such as avulsion and overall delta stability, are now better understood and the accuracy of the predictions made in this thesis can be tested for applicability in real systems. Furthermore, the morphodynamic framework advocated in this thesis provides a systematic, process-based understanding of the evolution of bifurcating deltas, which implicitly prescribes the evolution of the sediments that results from these processes and their architecture. This morphodynamic sedimentological model may be more illustrative than previous facies models, but it has yet to be tested.

5.2 Directions for future work

The preceding work has focused on the small-scale to reach-scale dynamics of delta channel networks. Understanding these local dynamics has uncovered many interesting directions of future research for understanding their importance at the larger scale. Firstly, I documented that the process of deltaic avulsion in flume experiments is caused by downstream growth of a river mouth bar. Future numerical and flume studies should address whether these avulsions operate in all parts of parameter space or if they are limited to certain conditions. Robust field documentation of this avulsion process and its linkage to mouth bar growth would be a substantial step forward in developing a universally accepted model of delta evolution.
Results show that equilibrium bifurcations distribute water and sediment asymmetrically, yet most evidence suggests deltas prograde symmetrically. How can deltas prograde symmetrically if the bifurcations route water and sediment asymmetrically? A more sophisticated, coupled channel network and open ocean model could explore the feedbacks that operate to make symmetrical delta progradation possible. Furthermore, all previous bifurcation stability work has been for generic bifurcation configurations and theoretical. More theoretical studies that explore the range of known bifurcation configurations coupled with field-monitoring of bifurcations through multiple flooding cycles could help delineate the stability fields of natural bifurcations.

Future work should also address the extent to which delta channel networks are affected by climate change. Results in this thesis demonstrated the entire delta channel network can be affected by perturbations, such as the closing of one channel. A natural extension of those results is to explore specific scenarios associated with climate change to understand the sensitivity of delta channel networks.

An advanced delta evolution model that includes in-channel processes as well as bank erosion, floodplain processes, and open ocean processes may be essential for tackling the important future problems. For example, observational evidence shows that delta floodplains are littered with scars of abandoned channels and these channels could play a significant role in focusing the locations of future avulsions. Also, the forces of cohesiveness, which act to keep banks stable, may be a necessary condition in generating these remarkably stable channel patterns and further tests are needed to understand the contribution of cohesiveness.
REFERENCES CITED


Amorosi, A., and S. Milli (2001), Late Quaternary depositional architecture of Po and Tevere river deltas (Italy) and worldwide comparison with coeval deltaic successions, *Sedimentary Geology, 144*, 357-375.


Best, J. (2005), Kinematics, topology and significance of dune-related macroturbulence: some observations from the laboratory and field, Special Publications International Association Sedimentology, 35, 41-60.


Frings, R., and M. G. Kleinhans (2008), Complex variations in sediment transport at three large river bifurcations during discharge waves in the river Rhine, *Sedimentology*.


Karssenberg, D., and J. S. Bridge (2008), A three-dimensional numerical model of sediment transport, erosion and deposition within a network of channel belts, floodplain and hill slope: extrinsic and intrinsic controls on floodplain dynamics and alluvial architecture, *Sedimentology, 1717-1745*. 


Krone, R. B. (1962), Flume studies of the transport of sediment in estuarial shoaling processes Final Report1-41 pp, Hydraulic Engineering Laboratory and Sanitary Engineering Research Laboratory, University of California, Berkeley.


Van Maren, D. S. (2005), Barrier formation on an actively prograding delta system; the Red River Delta, Vietnam, Marine Geology, 224, 123-143.


Waldrop, W. R., and R. C. Farmer (1973), Three dimensional flow and sediment transport at river-mouths, 137 pp., Coastal Studies Institute, Louisiana State University, Baton Rouge, LA.


APPENDIX A: Notation

\( C \) Chezy roughness, \( L^{1/2} T^{-1} \)
\( C' \) nondimensional Chezy bed roughness
channel \( a \) the unbifurcated channel
channel \( b \) the bifurcate channel with the smaller discharge
channel \( c \) the bifurcate channel with the larger discharge
\( D \) channel depth, \( L \)
\( D_{50} \) median bed grain size, \( L \)
\( D_r \) average depth ratio; high discharge channel divided by low discharge channel
\( g \) acceleration due gravity, \( L T^{-2} \)
\( h_b, h_c \) water surface elevations at the downstream boundaries of channels \( b \) and \( c \), \( L \)
\( I \) cross levee impulse per unit area, \( M L^{-1} T^{-1} \)
\( L \) bifurcate channel length, \( L \)
\( L' \) bifurcate channel length relative to the channel width
\( M \) momentum flux of turbulent jet
\( Q \) water discharge, \( L^3 T^{-1} \)
\( Q_r \) water discharge ratio; high discharge channel divided by low discharge channel
\( q_s \) sediment transport rate per unit width, \( L^2 T^{-1} \)
\( Q_s \) sediment transport rate, \( L^3 T^{-1} \)
\( Q_{s*} \) sediment transport rate in a symmetrical solution, \( L^3 T^{-1} \)
\( Q_{s_{bedload}} \) sediment transport rate of the bedload fraction, \( L^3 T^{-1} \)
\( Q_{s_r} \) sediment flux ratio; high discharge channel divided by low discharge channel
\( Q_{s_T} \) transverse sediment flux at the bifurcation, \( L^3 T^{-1} \)

subscript

\( a, b, c \) refers to either channel \( a, b, \) or \( c \)
\( t \) time, \( T \)
\( t_a \) time of avulsion, \( T \)
\( t_b \) time of river mouth bar stagnation, \( T \)
\( T_m \) morphological time scale, \( T \)
\( T_{ND} \) nondimensional time
\( V \) volume of sediment deposited during one time step, \( L^3 \)
\( V_a \) volume of sediment deposited from RMB stagnation to avulsion, \( L^3 \)
\( W \) channel top-width, \( L \)
\( W_o \) Channel top width of the zeroth order bifurcation, \( L \)
\( x, y \) planform dimension, \( L \)
\( \Delta E \) difference in bed elevation over one time step, \( L \)
\( \Theta \) Shields stress
\( \Theta_{crit} \) critical Shields stress for incipient motion of a given sediment size
\( \alpha \) channel aspect ratio; width divided by depth
\( \eta \) bed ramp height, \( L \)
\( \rho \) fluid density, \( M L^{-3} \)
\( \rho_s \) sediment density, \( M L^{-3} \)
\( \tau_o \) basal fluid shear stress \( M L^{-1} T^{-2} \)
APPENDIX A: Field Data

Appendix A contains additional data collected from the Mossy delta during the summers of 2006 and 2007. The data are presented in figures A1 through A-12 and consist of bathymetry, bed grain size and composition, suspended sediment concentrations, and water surface elevation data. Bathymetry data were collected with a single beam echo sounder and then were imported into Surfer v 7.0 and interpolated with a Kriging scheme. Bed grain size data were collected by dragging a weighted bucket on the bed of the river and therefore represent an integration of the first 5 to 10 centimeters. Those data were sieved and analyzed at Binghamton University by David Janesko. Suspended sediment concentration data were collected with a standard USGS sampler and were also analyzed at Binghamton. Water surface elevation data were collected with a differential global positioning system (see section 3.6 for more details). The data were collected over multiple days during the falling limb of a flood and were transformed to a common datum.
Figure A-1 Composite aerial photograph of Mossy Delta, Saskatchewan, Canada from 2003. Individual photos are from Information Services Corporation [in press]. The white line on the east side of the delta is the shoreline. The numbers mark the locations of the nine bifurcations for which I collected data.
Figure A-2  Data collected in 2006 for bifurcation 1 (see Figure A-1 for location).
Figure A-3  Data collected in 2007 for bifurcation 1 (see Figure A-1 for location).
Figure A-4 Data collected in 2006 for bifurcation 2 (see Figure A-1 for location).
Figure A-5  Data collected in 2006 for bifurcation 3 (see Figure A-1 for location).
Figure A-6  Data collected in 2007 for bifurcation 3 (see Figure A-1 for location).
Figure A-7  Data collected in 2006 for bifurcation 4 (see Figure A-1 for location).
Figure A-8  Data collected in 2006 for bifurcation 5 (see Figure A-1 for location).
Figure A-9  Data collected in 2007 for bifurcation 6 (see Figure A-1 for location).
Figure A-10  Data collected in 2007 for bifurcation 7 (see Figure A-1 for location).
Figure A-11  Data collected in 2006 for bifurcation 9 (see Figure A-1 for location).
Figure A-12  Composite water surface elevation map from the Mossy delta, Saskatchewan, Canada. Elevations are measured in meters above sea level.
APPENDIX B: Error Propagation in determining I

Appendix B contains five figures that show the error associated with each measurement of I, the cross-levee impulse, described and presented in Chapter 2. The error propagation was estimated using methods described in [Taylor, 1997]. Two variables, depth ($h$) and slope ($S$) in $I$ are subject to error. To calculate error propagation, the standard deviations of the slope and depth measurements are converted into fractional errors, which is the standard deviation divided by the mean, and then added together and multiplied by the best estimate of $I$. The fractional errors of slope and depth are calculated in a Matlab script that is available upon request.
Figure B-1  Estimated error in $I$ for DL2.
Figure B-2 Estimated error in I for DL4.
Figure B-3 Estimated error in $I$ for DL5.
Figure B-4  Estimated error in $I$ for DL9.
Figure B-5 Estimated error in $I$ for DL12.
VITA

Douglas A. Edmonds

EDUCATION:
M.Sc., The Pennsylvania State University
Department of Geosciences, spring 2006
Advisor: Dr. Rudy Slingerland

B.Sc., Saint Louis University, summa cum laude
Department of Earth and Atmospheric Sciences, 2003

RESEARCH EXPERIENCE:
2004-2009 Research Assistant, Penn State University, various NSF and PRF supported projects on deltas and their bifurcations

2006-2007 Research Geologist, Exxon Mobil Upstream Research, Reservoir Characterization Group

2005 Production Geologist, BP, Nakika Group, Deepwater Gulf of Mexico

2004 Consulting Geologist, State College, PA, Contact: Dr. Rudy Slingerland

TEACHING EXPERIENCE AT PENN STATE:
GEOSC 496, Field Camp, Summer 2004, 2007

GEOSC 340, Geomorphology, Fall 2006

GEOSC 310, History of the Earth, Spring 2006

PUBLICATIONS


Edmonds, D. A., Hoyal, DCJD, Sheets BA, and R. L. Slingerland, Predicting delta avulsions: Implications for coastal wetland restoration, in press at Geology