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Linking Form and Process in Braided Rivers Using Physical and Numerical Models

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LINKING FORM AND PROCESS IN BRAIDED RIVERS
USING PHYSICAL AND NUMERICAL MODELS

by

Alan Kasprak

A dissertation submitted in partial fulfillment
of the requirements for the degree
of

DOCTOR OF PHILOSOPHY

in

Watershed Sciences

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2015
ABSTRACT

Linking Form and Process in Braided Rivers
Using Physical and Numerical Models

by

Alan Kasprak, Doctor of Philosophy
Utah State University, 2015

Major Professor: Dr. Joseph M. Wheaton
Department: Watershed Sciences

Braided channels arise due to high sediment availability in conjunction with regular competent flows and readily erodible banks. Together, these boundary conditions lead to the deposition and reworking of a network of transient bars that characterize the braided planform. However, quantifying the geomorphic response of braided systems to alterations in these boundary conditions is not straightforward, as channels adjust over a wide range of timescales, rendering traditional field-based observation intractable. As such, the development of simple yet robust relationships between channel morphology and sediment transport has the potential to allow predictions of channel response to altered hydrologic or sediment regimes. In this research, I first use laboratory flume experiments to relate particle travel distance during floods (termed particle path length) and the spacing of channel bars in braided rivers (Chapter 2), finding that deposition sites for sediment in transport can be readily predicted by the characteristic confluence-diffuence spacing in a reach. I then use the relationship between path length and channel morphology to build a simple, open-source morphodynamic model for braided rivers that computes sediment transport using path-length distributions derived from bar spacing (Chapter 3). I explore the validity of this model, specifically noting that its modular framework allows exploration of process representations in morphodynamic modeling in ways existing models do not. Finally, I employ the model to determine the role of sediment supply in braided channel bar morphodynamics.
(Chapter 4). Specifically, I address the relative roles of sediment sourced from upstream versus sediment sourced from within a braided reach in terms of channel morphodynamics at decadal timescales. This research demonstrates that simple scaling relationships, while necessarily imperfect, nevertheless provide insight into morphodynamic processes in braided rivers, while also allowing predictions of channel response to sediment or hydrologic forcing at the timescales of channel adjustment.

(172 pages)
Braided rivers are characterized by their dynamic nature, and are often significantly reshaped during each flood capable of transporting sediment. Over time, they adjust in response to the frequency and magnitude of floods, along with the amount of sediment available for bar building. Factors such as climate change, dam construction, or land use alteration that change the amount of sediment or water available to braided rivers may subsequently affect channel form. One avenue toward understanding braided channel evolution is to develop simple relationships between channel form and sediment transport, and extrapolate those relationships over extended timescales. With funding from the National Science Foundation ($271,000), I first conducted laboratory experiments that linked the travel distance of sediment during a flood (termed \textit{particle path length}) and the spacing of bars in braided rivers. I then developed a simple model that predicts channel response to floods by transporting sediment according to specified path lengths. Finally, I employed the model to answer questions regarding the source of sediment used for bar building in braided rivers.

This research provides an important step in linking channel form and sediment transport in gravel-bed braided rivers, although the relationships developed here certainly deserve further testing across a variety of rivers and over floods of varying magnitude and duration. The predictive model developed herein provides a novel method for simulating channel evolution using a simple sediment transport approach. Additionally, the model is built using a modular framework that allows users to easily explore the effect of altering the way processes are represented, or whether they are included at all, on channel evolution.
ACKNOWLEDGMENTS

It’s an understatement to say that a Ph.D. takes a long time, and through the course of five years at Utah State, I’ve been fortunate to meet people who have helped me along the way more than I can put into words. Some people, though, have been by my side for much longer than that. My family has always encouraged me to chase my dreams and put in my best effort; that I’m able to finish this degree is a reflection of them teaching me to be hard-working and perseverant for as long as I can remember.

My advisor at Utah State, Joe Wheaton, has been as much a colleague as a supervisor, and his willingness to offer scientific guidance as well as his flexibility to let me pursue my own ideas and project directions are greatly appreciated. The members of my committee, Nick Bouwes, Patrick Belmont, Joel Pederson, and Jack Schmidt, have improved my work immeasurably through many conversations, a great deal of constructive feedback on my writing, and yes - even by challenging me during my comprehensive exams.

Funding for my work was provided by the National Science Foundation, with additional support from Utah State University. The Community Surface Dynamics Modeling System and National Center for Earth Surface Dynamics contributed laboratory time and computational resources. Additionally, Peter Ashmore and Sarah Peirce donated laboratory facilities and their time and intellect to make my work at the University of Western Ontario possible.

I’m indebted to the past and present members of the Ecogeomorphic and Topographic Analysis Laboratory (ET-AL) and associated labs at Utah State, especially Sara Bangen, Reid Camp, Florie Consolati, Kenny DeMeurichy, James Hensleigh, Nate Hough-Snee, Wally MacFarlane, Elijah Portugal, William Leonard Roberts, and Justin Stout.

The morphodynamic modeling component of this work wouldn’t have been possible without Konrad Hafen, who put aside an undergraduate degree in fisheries to learn C++ from scratch, spending hours writing the lines of code that made this project happen. Philip Bailey of North Arrow Research and Matt Nahorniak of South Fork Research were always on call for technical support and coding advice.
A number of generous researchers assisted with field work in Scotland and took the time to discuss my work while I was in the UK, including James Brasington, Niall Lehane, Julian Leyland, David Sear, Mark Smith, and Richard Williams.

In the last year of my Ph.D., I’ve been based in Flagstaff, Arizona. The scientists and staff at the U.S. Geological Survey’s Southwest Biological Science Center and Grand Canyon Monitoring and Research Center, in particular Paul Grams, Dave Lytle, and Joel Sankey have generously provided me with the best office I’ve ever had, along with the resources to help me stay productive away from Logan.

Finally, I was fortunate to spend my time in Logan with a group of people that made five years of school pass by more quickly than I could have imagined. Jake, Natalie, Nick, Schroer, Tony, Harrison, Stephen, Hannah, Dave, Bowie, and EWall - thanks for everything. And Lisa, thanks for always finding a way to make me smile, no matter the crisis-of-the-day. There’s nothing I could possibly write here that would come close to letting you know how much you helped along the way, but just know I’m lucky to have had you by my side and I couldn’t have done it without you.

Alan Kasprak
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INTRODUCTION

1.1 Motivation and Background

Braided channels result from high sediment availability coupled with regularly-occurring competent flows and readily erodible banks [Ashmore, 1991; Leopold and Wolman, 1957]. Together, these boundary conditions give rise to a dynamic network of channels splitting and joining around mid-channel and bank-attached bars [Knighton, 1998]. It has been argued that braided rivers represent the 'default' channel state [Lane, 2006], and laboratory experiments and the geologic record indicate that single-thread channels are largely the result of bank cohesion offered by the presence of vegetation [Tal and Paola, 2007].

The reasons channels may take a braided planform as opposed to a single-thread pattern have been studied by numerous researchers. Sediment supply, bank stability, channel slope, and the ability to mobilize bed material have all been pinpointed as factors that lead to the existence of braided channels [Leopold and Wolman, 1957; Gran and Paola, 2001; Tal and Paola, 2007; Braudrick et al., 2009; Mueller and Pitlick, 2014]. There are numerous cases in the literature where braided channels transitioned to single-thread systems, or vice versa, and these are typically attributed to shifts in climate that drive altered sediment or hydrologic regimes, or alternatively changes in land use or river regulation [Gurnell et al., 2009].

Predicting the response of channels to altered sediment supply and hydrology is not straightforward. This is especially true in braided channels, where reach-scale morphology may be completely reshaped during the course of a single flood, while true planform shifts may manifest over years to centuries (e.g., dynamism versus persistence; [Czarnomski et al., 2012; Wheaton et al., 2013]). The timescales of these changes render traditional field-based observation intractable for quantifying channel evolution (see Chapter 2 and [Kasprak et al., 2015]).

An alternative to field-based techniques for tracking braided channel evolution is the development of simple relationships that link channel form and sediment transport, which can
be extrapolated to understand how altered sediment supply or hydrology manifest as morphologic changes over extended spatiotemporal scales. For example, Schmidt and Wilcock [2008] and Grant [2003] employed simple indices describing alterations in flood regime and sediment by dams to forecast geomorphic response of downstream channel reaches. In another example, Pyrce and Ashmore [2003a] used marked tracer particles in a meandering single-thread laboratory channel to document the relationship between particle travel distance during floods, or path length, and the spacing of downstream point bars. In braided rivers, Hundey and Ashmore [2009] found consistent geometric relationships between anabranch width and confluence-diffuence spacing, suggesting that particle travel distances may correspond to the characteristic length scale at which anabranches join and split (e.g., the location of bars).

Previous research that has formed the foundation for relating channel form and sediment transport processes [e.g., Pyrce and Ashmore 2003a, 2003b; Hundey and Ashmore, 2009; Habersack, 2001; Mueller and Pitlick, 2014] has provided an important first step in linking form and process in gravel-bed braided rivers. Building on this work, this dissertation exclusively focuses on braided systems where the substrate is predominantly gravel-sized sediment. Although sand-bed braided rivers are common (see Bristow and Best, 1993), prior research that has investigated the relationship between channel form and sediment transport processes in gravel-bed braided rivers make these systems an ideal test case for exploration of form-process linkages in this work.

Throughout this dissertation, I use terms describing sediment in rivers that are deserving of explicit definition. I use the umbrella term 'sediment availability' to mean the entirety of sediment that may be entrained in a given reach of river; this term does not differentiate between 'local' sources of sediment and 'upstream' sources of sediment. In the case of the former, 'local' sediment supply is that volume of sediment available for entrainment, transport, and deposition that is located within a given river reach (e.g., found within bars or in the channel bed/subsurface). In contrast 'upstream' sediment is that volume of sediment that is imported into a given reach from upstream. Note that these terms are
time-dependent: 'upstream' sediment may become 'local' sediment over a given time period if it is imported into the reach, deposited there, and subsequently reworked by the channel.

1.2 Dissertation Objectives

The objective of this work is to explore the predictability of sediment transport distances as a function of bar spacing, and the utility of that relationship as a tool to model braided river morphodynamics. Each of the three research projects detailed here contains objectives related to this overall goal. In Chapter 2, I use laboratory flume work to determine whether braided channel morphometry can act as a predictor of sediment travel distances during floods. In Chapter 3, I seek to understand the utility of that relationship as a foundation for morphodynamic modeling of braided rivers. Finally, in Chapter 4, I leverage this morphodynamic model to explore the effect of several sediment supply scenarios on braided channel evolution at decadal scales.

1.3 Organization of the Dissertation

In this research, I investigate whether channel morphology and sediment path lengths are related in braided rivers, and subsequently develop a simple model that transports sediment according to morphologic unit spacing. Ultimately, I employ the model to answer basic questions about the importance of sediment source and supply on bar morphodynamics in braided rivers.

Chapter 2 investigates the relationship between particle path lengths and the location/spacing of channel bars in braided rivers. Previous work in single-thread channels has indicated particle deposition sites may correspond to the location of bars encountered by sediment in transit [Pyrce and Ashmore, 2003a, 2003b]. In the summer of 2013, I traveled to the University of Western Ontario and conducted a series of five experiments using a laboratory model of a gravel-bed braided river. I examined intra-flood morphodynamics using high-resolution digital elevation model construction and differencing [Wheaton et al., 2010], and tracked particle travel distances using fluorescent tracer particles and visual
tracer recovery. For all experiments, I constructed particle path length distributions and analyzed the relationship between the location of in-channel morphologic units (e.g., bars) and particle deposition locations.

Chapter 3 leverages the findings from my laboratory flume experiments to develop a numerical model that predicts braided river evolution. Previous efforts aimed at morphodynamic modeling of braided rivers have generally taken one of two approaches: employing reduced complexity physics to capture the relevant spatiotemporal scales of channel evolution, or using faithful representations of the relevant physics to accurately model hydraulics and sediment transport. In the case of the former, the inability to conserve momentum leads to large-scale model inaccuracies, particularly with regard to hydraulic modeling [Coulthard et al., 2007]. With regard to the latter, the computational overhead required to obtain solutions restricts models to very fine spatiotemporal scales (e.g., meter-scale, events to months; [Brasington and Richards, 2007]). The model I develop in Chapter 3 employs highly simplified sediment transport routines to mobilize volumes of material downstream according to morphologically-based path-length distributions. I compare the results of morphodynamic modeling with field-based surveys on two rivers at timescales ranging from single floods to a decade.

Chapter 4 employs the morphodynamic model developed in Chapter 3 to explore the influence of sediment source on bar morphodynamics in braided rivers at decadal timescales. It is widely accepted that high rates of sediment supply are a necessary precondition for braided planform development and persistence [Ashmore, 1991]. However, it remains unclear whether this sediment must necessarily be supplied from upstream, or alternatively whether local sources can supply adequate sediment for braided planform maintenance. Braided rivers are often found in valley bottoms with abundant sediment sourced from tributary inputs or glacial deposits, for example [Miall, 1977; Bristow and Best, 1993]. As a result, it is possible that the sediment used to build and maintain bars may be largely sourced from within braided reaches. To disentangle the relative role of upstream versus local, within-reach sediment supply on bar morphodynamics, I employ scenario-based mor-
phodynamic modeling in concert with digital elevation model differencing and mechanistic segregation of sediment budgets [e.g., Wheaton et al., 2013] at decadal timescales on a braided gravel-bed river.

Chapter 5 details the general conclusions of this work and provides a synopsis of my findings. There are instances where this research would benefit from additional exploration and the development of an event-scale channel evolution framework in braided rivers. As a result, potential future directions of research are suggested for building on the findings of this dissertation.
References


CHAPTER 2

THE RELATIONSHIP BETWEEN PARTICLE TRAVEL DISTANCE AND CHANNEL MORPHOLOGY: RESULTS FROM PHYSICAL MODELS OF BRAIDED RIVERS

Abstract

Channel form and sediment transport are closely linked in alluvial rivers, and as such the development of a conceptual framework for the downstream controls on particle mobility and likely deposition sites has immense value in terms of the way we understand and predictively model rivers. Despite the development of conceptual models which frame flood-scale particle transport distance (termed path length) as a function of channel bar locations, an understanding of the controls on such path lengths in braided rivers remains especially elusive, in large part due to the difficulty in explicitly linking morphology and particle transport distances in the field. Here we utilize a series of laboratory flume experiments to link path length distances with channel morphology. Our morphologic characterization is based on ultra-high-resolution digital elevation models and bar classifications derived from structure-from-motion topography, while we simultaneously capture particle path lengths using fluorescent tracer particles over the course of five physical model simulations. Our findings underscore the importance of channel bars in acting as deposition sites for particles in transport; 81% of recovered tracers were found in association with compound, point, lateral, or diagonal bars. Bar heads (29%) and bar margins (41%) were the most common bar-related deposition surfaces for recovered tracers. Peaks in particle deposition frequency corresponding to channel bars were often noted on path-length distributions from tracer data; most tracers were deposited in areas that had experienced shallow ($\Delta z = 0.002$ m) deposition. Average path length distance (2.5 m) was closely related to average confluence-diffuence spacing (2.3 m) across all runs. The transferability of this understanding to

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1This chapter is published in *Journal of Geophysical Research: Earth Surface* and co-authored with Peter Ashmore, Sarah Peirce (University of Western Ontario) and James Hensleigh and Joseph Wheaton (Utah State University).
braided streams has important implications for the development of simplified morphodynamic models which seek to predict braided channel evolution across multi-flood timescales.

2.1 Introduction

In alluvial rivers, channel morphology and sediment transport processes are inextricably linked [Church, 2006]. Despite this form-process coupling arising as a result of the self-formed nature of the channel, studies that have explicitly attempted to document the downstream mobilization of particles and their deposition in association with geomorphic units are rare in gravel bed streams [Pyrce and Ashmore, 2003a]. While it has been surmised that particle path length (the distance a particle is transported downstream during the course of a single flood event; [Pyrce and Ashmore, 2003a, 2003b]) is influenced by channel morphology such as the location of pools, bars, or eddies, and conversely that path length must, through some feedback process, determine the initial spacing of channel bars, few studies have attempted to jointly quantify these path lengths and their association with channel form [Church, 2006]. Simple correlation efforts between channel physical attributes and downstream transport distances of particles in natural streams have noted an association between channel width and path length [Beechie, 2001] while hypothesizing that channel width may appear to drive particle travel distance insofar as it also drives bar spacing [Leopold and Wolman, 1957]. Although particle size may play a role in the downstream transport distance of particles at the coarser end-member of the bed grain size distribution (e.g., $D > 2 \text{D}_{50}$), it is likely that for intermediate grain sizes on the bed of gravel-bed rivers, transport distance may well be independent of clast diameter [Church and Hassan, 1992]. Rather, particle transport distance may instead reflect the efficacy of channel morphology in promoting deposition sites for sediment in transit, as full mobility implies no influence of particle size on path length, except for local sorting effects at channel bends and bars [e.g., Wilcock, 1997].

While studies that have employed tracers to document particle mobility during high flows are common in gravel bed rivers [see Lamarre, 2005], relating sediment transport to
channel morphology at the event scale is difficult because it requires a complete topographic survey of the channel reach in question both prior to and following a flood. Although the advent of rapid surveying techniques (e.g., terrestrial laser scanning; [Williams et al., 2013]) has made this problem more tractable, much of our understanding of morphologic controls on particle path length has necessarily been the result of laboratory flume studies, which have revealed an apparent coupling between channel units and particle deposition. In single-thread channels, particles sourced from areas upstream of bends appear to be most frequently deposited on downstream point bars [Pyrce and Ashmore, 2003a, 2003b], with the highest probability of deposition corresponding to the location of the first point bar encountered and subsequently decreasing at more distal point bars. Independent of channel morphology, flood magnitude may play a role in determining downstream mobility of particles during a transport event (here we refer to data describing the path lengths of multiple particles as a path length distribution, typically presented as a histogram; [Pyrce and Ashmore, 2003a, 2003b]). At low magnitude/low bed mobility floods, path length distributions appear to be positively skewed, with many particles exhibiting non-movement or movement over very short downstream distances. In this case, downstream channel bars may play a more pronounced role in trapping particles as flow and relative mobility increases up to the formative discharge of the channel.

We argue that the extension of such rule sets documenting the relationship between channel morphology and particle path length, which were initially developed in single-thread channels, may be of particular importance in braided rivers. In these settings, frequent discharge fluctuations, readily-erodible banks, and a high degree of bedload mobility lead to extremely dynamic channel planforms with frequent downstream mobilization of bed sediment (gravel particles) through a continually-shifting network of mid-channel and bank-attached bars that may act as deposition sites for particles in transport [Ashmore, 2013]. While this combination of factors makes braided rivers ideal sites in which to observe the relationships between particle path lengths and channel morphology, comparatively few studies have done so, either in braided river flumes or natural channels. Existing studies have
hypothesized the potential importance of a path length/morphologic coupling whereby bar spacing truncates downstream sediment transport distances, promoting a positive feedback acting to reinforce the location and persistence of channel bars [Hundey and Ashmore, 2009]. Indeed, the earlier work of Habersack [2001] provided evidence for this hypothesis, employing radio tracking to document the preferential deposition of several mobilized particles on an aggrading bar in a New Zealand braided river. Given the high-resolution radio tracking of particles by Habersack [2001], sample size was limited (n = 16) and the degree to which the tracked particles were representative of overall reach-scale sediment transport distances is difficult to discern.

Despite this previous research, a critical knowledge gap remains in our ability to connect channel morphology and path lengths in braided streams [Church, 2006] - largely the result of insufficient data concurrently linking particle transport and channel morphologic evolution. This study seeks to close that knowledge gap with a series of physical experiments. Here we use laboratory flume studies of braided rivers, in concert with fluorescent tracer particles recovered following experimental floods via high-resolution digital photography to document downstream particle mobilization. Along with these data on downstream particle mobility, we employ structure-from-motion photogrammetric techniques to derive millimeter-scale topographic datasets describing channel morphology and subsequently delineate geomorphic units where tracer particles are deposited. Finally, by differencing and mechanistically-segregating these elevation datasets before and following floods, we gain a greater understanding of the role of sediment budget imbalance and braiding mechanisms in driving particle deposition in various geomorphic units [Ashmore, 1991; Ashmore, 2013; Wheaton et al., 2013].

2.2 Methods

The experiments conducted during this study took place at an indoor flume facility at the University of Western Ontario (UWO). Here we describe (a) the physical characteristics of the flume and the experimental flows, (b) the generation of digital elevation models
(DEMs) from the flume surface, (c) differencing of DEMs and mechanistic segregation of morphologic change to relate sediment transport and channel form during runs, and (d) sediment tracer particles used in the experiments, along with their seeding and recovery. In all, five separate 'floods' or runs were conducted using this flume, and the workflow for each run is shown in Figure 2.1. Each step within this workflow is described in greater detail in the subsections that follow.

2.2.1 Flume

The flume at UWO measures 20 m long by 3 m wide, and is an approximately 1:35 Froude-scaled model of the gravel-bed braided Sunwapta River in Alberta, Canada. The mechanics behind Froude-scale modeling are well established, and have been discussed in
previous studies and in work using this same flume [see Peakall et al., 1996; Warburton et al., 1996; Egozi and Ashmore, 2009; Gardner and Ashmore, 2011]; in brief, Froude-scaled models preserve non-dimensional bed shear stress (Shields stress) such that near-turbulent flow, bed roughness, and resultant sediment transport conditions are similar to those found in natural channels. All flume runs were conducted at a constant bed slope of 1.5%. The bed of the flume is composed of grains with a $D_{50}$ of 1.2 mm and a $D_{90}$ of 3.4 mm, corresponding to coarse gravel when upscaled by a factor of 35 [Ashmore et al., 2011]. A steady water discharge of 2 $\text{l s}^{-1}$ was used for all flume runs, which is approximately equivalent to 14 $\text{m}^3\text{s}^{-1}$ (using the 1:35 length scaling; the degree to which time exhibits a similar scaling is unknown). This discharge corresponds to the formative flood of the braided channel which was evolved from a plane bed for $\sim$12 hours prior to the beginning of experiments, whereby an initially straight channel carved into the plane bed grew more and more sinuous until cutoffs and avulsions became commonplace and a braided channel planform was developed.

Across all runs, the average braidplain width was 1.7 m and the average main channel width was 0.44 m.

2.2.2 Structure-from-Motion DEMs

The UWO flume is outfitted with a camera platform mounted on rails 2.9 m above the flume bed, allowing for capture of overhead photographs. For this purpose, we used a Canon 10D digital SLR camera with a zoom lens fixed at a focal length of 20 mm. Photograph sets were captured over the dry flume bed both before and following each flume run (Figure 2.1). Using a previously-surveyed network of ground control targets for geo-referencing in conjunction with the software package Agisoft, we merged sets of overhead photographs to create a continuous down-flume photograph mosaic. Once stitched and orthorectified, the photograph mosaics had a pixel resolution of 0.001 m.

Using the structure-from-motion (SfM) component of Agisoft, we created point clouds and then used the built-in point-to-surface interpolation feature in Agisoft to produce seamless digital elevation models (DEMs) from these point clouds. SfM DEM generation required
accurate lens calibration to account and correct for inherent radial distortion in the photographs. This calibration was completed using the freely-available Agisoft Lens software. Over all runs, point clouds contained on average $2.5 \times 10^6$ points. We used a DEM resolution of 0.003 m for consistency with the DEMs produced via traditional photogrammetric techniques during previous work using this flume \cite{Gardner and Ashmore, 2011}, and this resolution is similar to the grain roughness of the bed material.

\textit{Gardner and Ashmore} \cite{2011} estimated the vertical accuracy of DEMs created from stereo photogrammetric techniques using this flume at 0.002 m. As an independent assessment of vertical DEM accuracy, we examined coincident points (defined here as individual points lying within 0.0001 m of one another) within the dense SfM point cloud. Where such coincident points existed, the difference of the points' elevation was computed to assess the relative accuracy of the DEM. Differences of all coincident points resulted in a mean of 0.001 m $\pm$ 0.002 m, indicating our DEMs had internal errors that were consistent with those DEMs produced by \textit{Gardner and Ashmore} \cite{2011} via traditional photogrammetric techniques.

2.2.3 DEM Differencing and Budget Segregation

To quantify geomorphic change during each flume run, we differenced DEMs using the Geomorphic Change Detection 5.2 software (GCD; \url{http://gcd.joewheaton.org}). This technique has been widely applied in studies investigating changes in channel morphology and sediment budgeting \cite{Wheaton et al., 2010; Grams et al., 2013}, but at its core involves the subtraction of an original DEM created from data surveyed prior to changes occurring from a newer DEM which was generated from data collected following geomorphic change. This differencing creates a DEM-of-Difference, or DoD, that describes elevation changes which have occurred during the period between topographic surveys, in turn yielding the change in volumetric sediment storage that occurred during the inter-survey period.

One of the most challenging aspects of analyzing a DoD is the fact that each dataset used in its generation contains an inherent level of error or uncertainty, which may then
be propagated into the resulting DoD. The methods behind quantifying and accounting for these errors have been studied in detail by Wheaton et al. [2010], and the GCD software employed in our analyses contains routines for their computation. Because DEM errors are spatially variable and dependent on multiple characteristics of the survey surface [e.g., Wheaton et al., 2010], our flume-based DEMs are prime candidates for differencing analysis using a fuzzy inference system (FIS), which produces estimates of resultant surface uncertainty from multiple datasets describing the factors that may influence that uncertainty. In short, FIS implementation requires (a) datasets describing each of the surface metrics contributing to DEM uncertainty, along with (b) a derived relationship between these surface metrics and elevation uncertainty. We produced slope rasters for each survey using ArcGIS (ESRI, Redlands, CA), and additionally decimated the dense SfM point cloud using the Topographic Point Cloud Analysis Toolkit (ToPCAT; Brasington et al., 2012) and produced surface roughness rasters for each DEM (Figure 2.2). To assess the effect of slope and surface roughness on DEM accuracy, we again used the previously-described coincident point analysis, and as we also knew the slope and surface roughness values at these coincident points, we could thereby assess the influence of each on resultant point-based DEM accuracy. Both slope (F-statistic = 30,342 at 1,711,534 degrees of freedom) and surface roughness (F-statistic = 1,354,052 at 1,711,534 degrees of freedom) were significantly correlated with elevation uncertainty.

The range of slope and roughness values across all surveys were divided into three categories: low (values less than the mean minus two standard deviations), medium (values falling between the mean ± two standard deviations), and high (values greater than the mean plus two standard deviations). The groupings of slope and roughness were combined using the FIS to yield resultant elevation uncertainties (low, moderate, high, extreme); the combination of slope and roughness and the resultant elevation uncertainties are shown in the rule sets in Figure 2.2. The power of an FIS lies in its ability to rapidly translate the effects of several input variables (here surface slope and roughness) into a resultant output value. In our case there is a large degree of uncertainty regarding the exact effect of each
input variable on DEM uncertainty, and this underscores the value of being able to use adjective-based rulesets to describe these effects (e.g., 'if slope is high and roughness is high then elevation uncertainty is extreme'). The output of the FIS was an estimate of DEM vertical uncertainty on a pixel-by-pixel basis, in effect a translation of an adjective-based fuzzy output (e.g., 'high uncertainty') to a crisp numerically-based uncertainty (e.g., 'an uncertainty of 0.02 m').

This still left the task of differencing the two DEMs and determining the resultant...
(propagated) uncertainty in each pixel of the DoD. To do this, we employed traditional propagation of error [Lane et al., 2003; Brasington et al., 2003], whereby the resultant vertical uncertainty in each pixel of the DoD ($U_{DoD}$) is computed as square root of the sum of squares of the individual pixel uncertainties used in the differencing ($U_{DEM1}$, $U_{DEM2}$), such that:

$$U_{DoD} = \sqrt{U_{DEM1}^2 + U_{DEM2}^2}$$

(2.1)

Once the propagated uncertainty was known, we employed probabilistic thresholding to each DoD so as to retain only those changes in our final DoD that could be reliably distinguished from changes arising due to survey noise in the constituent DEMs. The probabilistic thresholding method is described further by Brasington et al. [2003], but in short allows for thresholding of changes in a DoD as a function of a user-specified confidence in our ability to distinguish them from changes arising from survey errors. By computing the ratio of change ($Z_{DEM\,NEW} - Z_{DEM\,OLD}$) to the minimum level of detection ($minLoD$) on a pixel-by-pixel basis across a DoD (in effect a $t$-test; Equation 2), we can estimate a level of confidence for an elevation change of a particular magnitude.

$$t = \frac{|Z_{DEM\,NEW} - Z_{DEM\,OLD}|}{minLoD}$$

(2.2)

Here we used a confidence level of 99%, as we found this high confidence level, while conservative, most reliably discriminated between areas of true change and areas we knew showed erroneous changes arising from survey noise (e.g., steeply-sloping walls of the flume) better than a simple 95% confidence level; Figure 2.2 gives an example of this probabilistic thresholding as applied to DEMs generated via SfM.

DoDs were used both for the computation of the change in sediment storage during each flume run, and also for identifying braiding mechanisms and other fluvial processes using budget segregation of volumetric change. Wheaton et al. [2013] detailed how differences in braiding mechanism (in effect, those processes which act to promote and reinforce the
braided channel planform) vary widely on annual timescales and may vary in response to the sediment budget and/or hydrologic regime of a survey reach. Here we sought to understand the role of individual braiding mechanisms in promoting tracer deposition - insofar as they may produce channel forms which acted as preferential deposition sites for particles. Ashmore [1991] documented four primary braiding mechanisms (central bar development, chute cutoff of point bars, lobe dissection, and transverse bar conversion), to which Wheaton et al. [2013] added an additional seven mechanisms not unique to braided rivers (bar sculpting, channel incision, bank erosion, overbank sheets, confluence pool scour, lateral bar development, questionable or unresolved changes). The dynamics and distinguishing features of each of these braiding mechanisms are detailed both by Ashmore [1991] and Wheaton [2013]. We segregated areas of change in each DoD in terms of the braiding mechanism we could best infer as being responsible for those changes; we subsequently used the budget segregation feature in GCD 5.2 [Wheaton et al., 2010] to quantify the contribution of each braiding mechanism to the total volumetric change occurring in each DoD. Note that this volumetric contribution metric differs from areal extent of a particular braiding mechanism because the volumetric contribution of a particular mechanism may be equal to zero even though certain geomorphic features resulted from that mechanism during a flume run, if the erosional and depositional signatures of that mechanism were equal (e.g., no net volumetric change).

2.2.4 Tracers

In each run, we seeded and recovered three colors of fluorescent tracer particles: green, orange and blue, each with $D_{50} = 2.4$ mm. One hundred particles of each color were seeded during each run. In our fifth and final flume run, we also employed yellow tracers; fifty of these tracers were used in this run. As seeding prior to beginning water flow would result in many tracers being washed downstream with the first wave of water passing their location, we instead initiated water flow and waited until the flow was generally confined to the channels on the flume bed rather than consisting mainly of sheet wash over the
braid plain (approximately 5 minutes). At this point, tracers were injected into the flow at a particular morphologic location. Although this injection technique may have affected the entrainment of tracers, in this research we were mainly concerned with their ultimate depositional location, as opposed to studying their mobilization. Tracer seed locations are discussed within the results for each flume run (Section 2.3), but we generally tried to vary the locations of tracers as much as possible from run-to-run in order to assess the effect of seed location on depositional location (and hence the role of channel morphology on influencing sediment path lengths). To this end, the use of multiple tracer colors in each run allowed us to examine the effect of particle source location on resultant deposition location.

Tracers were recovered visually by marking their locations on geo-referenced post-flood photograph mosaics in ArcGIS (ESRI, Redlands, CA). For consistency of recovery effort, photographs were examined for two hours; for all mosaics, the rate of tracer recovery had slowed considerably after two hours of examination. Tracer seed locations were marked using high resolution overhead videos of flume runs in concert with the ground control targets located in the flume. In several instances, a large non-movement mode of tracers resulted in a patch of particles remaining at the seed location, thereby allowing easier delineation of the injection point. Tracer paths were inferred by constructing polylines in ArcGIS along channel anabranches from the tracer seed location to its recovery location. Where tracers were deposited on bars (either mid-channel or bank attached), we devised a simple bar surface classification scheme and further noted the location of these tracers on bars. Bars were divided into bar heads (the upstream edge of bars, typically widening in a downstream direction and found at the diffuence of two anabranches), bar tails (the downstream edge of bars, typically narrowing in a downstream direction and found at the confluence of two anabranches), bar tops (the main bar surface between the bar heads and bar tails), bar margins (the ~10 cm-wide 'fringes' of the main bar surface found between the bar heads and bar tails and distinguished from bar tops by their sloped profile extending down to the channel bed), and bar chutes which were dissecting the tops of bars. Note that
Figure 2.3. Bar Surface Delineation. Shown are bar heads, margins, tops, and tails in a point bar and a mid-channel compound bar. Bar chutes are also delineated dissecting the surface of the compound bar. Base is a hillshaded 0.003 m-resolution DEM.

Here we differentiate between chute cutoff (a braiding mechanism) and bar chutes, which were small headward-migrating incision features across the tops of bar surfaces. A visual representation of these surfaces on both mid-channel and lateral bars is shown in Figure 2.3.

Because of the visual methods employed to recover tracers, any non-recovered tracers could potentially be the result of one or more of the following: (a) failure to locate and mark all visible tracers on photographs, (b) burial of tracers rendering them invisible on photographs, and/or (c) export of tracers from the flume. We argue that case (a) likely contributes little to low recovery rates, as two hours was more than sufficient for the rate of visual recovery to decrease considerably, and all parts of the photograph mosaics were searched multiple times throughout the course of the two-hour recovery time. However, cases (b) and (c) may significantly contribute to low recovery rates. Alternatively, erroneously high recovery rates may be due to the import of tracers that were exported from the flume and subsequently recirculated through the flume’s sediment feed. However, we attempted to minimize this effect as much as possible by removing all visible tracers following each run with the aid of an ultraviolet lamp (causing the tracers to fluoresce and become readily visible, even when buried to 1-2 cm). Ultraviolet-based tracer recovery has been used for quite high recovery rates (95-97%) in previous research [Pyrce and Ashmore, 2003b], although that study involved a single flume run followed by rigorous scour and deeper
disturbance of the bed for locating tracer particles, while in this study we traded potentially lower tracer recovery rates in favor of relative preservation of the channel bed between runs to allow for a greater number of individual runs.

2.3 Results

For each of five flume runs, tracer recovery rates varied between 22 and 45% of tracer particles, consistent with many field-based studies which have used visual tracer recovery [Ashworth and Ferguson, 1989; Schmidt and Ergenzinger, 1992; Hattingh and Illenberger, 1995; Sear, 1996]. For all analyses reported here, we restrict our sampling window to an area of interest marked by tracer mobility, extending from the tracer seed locations (upstream boundary) to the most downstream tracer particle that was recovered in each run. The sediment imbalance of this surveyed area of the flume during the runs was consistently aggradational (e.g., sediment deposition, on average 11% imbalance). The exception to this was Run 2, in which the flume lost 3% more sediment than was deposited. The differences in these sediment budgets depending on the flume run (and hence the region examined) are not surprising and previous research has underscored the effect of sampling scale on both sign and magnitude of sediment budgets [see Grams et al., 2013], along with the rapid spatiotemporal variation of sediment flux in braided rivers [Ashmore et al., 2011].

For particles that moved, the most common depositional sites for tracer particles were compound bars (68% of recovered tracers), anabranches (12%) and point bars (9%). Across all five runs, the most volumetrically-significant braiding mechanisms were the development of central bars (34% of total change on average) and lateral bars (25%), along with bank erosion (22%). Of the bar surfaces delineated (bar heads, tails, tops, chutes, and margins), 70% of recovered tracers found on bars were either found on the heads of bars or the margins of bars, with the remaining tracers (30%) being recovered from bar tops and chutes dissecting bars. 3% of all recovered tracers were found on the tails of bars.

This section details the results of each flume run with respect to (a) sediment budgeting and morphodynamic change, (b) braiding mechanisms, and (c) the influence of channel form
on tracer deposition locations. A summary of results from all five runs can be found in Table 2.1.
Table 2.1. Sediment and Tracer Data for All Flume Runs

<table>
<thead>
<tr>
<th>Run Number</th>
<th>$V_{\text{Deposition}}$ (cm$^3$)</th>
<th>$V_{\text{Erosion}}$ (cm$^3$)</th>
<th>$V_{\text{Difference}}$ (cm$^3$)</th>
<th>% Imbalance</th>
<th>Recovery Rate (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>4725</td>
<td>4316</td>
<td>+409</td>
<td>+2</td>
<td>22</td>
</tr>
<tr>
<td>2</td>
<td>2484</td>
<td>2750</td>
<td>-266</td>
<td>-3</td>
<td>26</td>
</tr>
<tr>
<td>3</td>
<td>3247</td>
<td>1145</td>
<td>+2102</td>
<td>+24</td>
<td>36</td>
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<tr>
<td>4</td>
<td>4652</td>
<td>2161</td>
<td>+2491</td>
<td>+18</td>
<td>33</td>
</tr>
<tr>
<td>5</td>
<td>6746</td>
<td>3924</td>
<td>+2822</td>
<td>+13</td>
<td>45</td>
</tr>
</tbody>
</table>
2.3.1 Run 1

This run was slightly net depositional overall, with a net volumetric sediment imbalance of +2% (Figure 2.4). A combination of several braiding mechanisms contributed nearly equally to morphologic changes in this run, including bank erosion, central and lateral bar development, chute cutoff, and channel incision (Figure 2.4). DoDs and mechanisms of change are shown in Figure 2.4. This first run was highlighted by the erosion of channel banks and scour of developing anabranches around a relatively un-dissected mid-channel bar that was a remnant of the original plane-bed morphology of the flume. Tracers were seeded in two anabranches upstream of a confluence scour pool (Figure 2.4). In general, tracers were mobilized downstream with few particles remaining at the seed location (Figure 2.4); tracer recovery rate was 22%. Irrespective of seed location, tracers were generally recovered 1.5 - 3 m downstream, with the majority of tracer particles being deposited on the head and river-right margin of a compound bar, while fewer tracers were deposited downstream atop a diagonal bar surface. For this and all subsequent tracer travel distance plots, downstream transport distances were referenced to the seeding location of the most upstream tracer color to allow for comparison of recovery locations using common downflume coordinates. Relatively few particles were deposited along the river-left margin of the first downstream compound bar surface.

2.3.2 Run 2

Run 2 was the only net erosional run (i.e., net loss of sediment from the reach) in terms of the sediment budget, with a loss of -3% (Figure 2.5). The most volumetrically-significant braiding mechanisms were bank erosion (43%) and the development of lateral (29%) and central bars (14%). Large swaths of eroding banks, deposition of eroded material along the tops and margins of mid-channel bars, and the large-scale dissection of the mid-channel island seen in Run 1 by head-cutting chutes were seen in Run 2. Tracer particles in this run were again seeded in anabranches upstream of a confluence pool. Recovery rate was 26%, and tracers were predominantly found 0.5 - 3 m downstream, with sub-peaks in depositional
frequency corresponding to the head of the large mid-channel compound bar immediately
downstream of the seed location, along with tracers that were deposited on both the river
left and river right margins of this compound bar (Figure 2.5). The remaining tracers were
most often recovered from within head-cutting chutes dissecting the top of the compound
bar.

2.3.3 Run 3

Run 3 was net depositional, with a net sediment imbalance of +24% over the area
of interest (Figure 2.6). The development of central bars (50%) along with lateral bar
development (25%) and bank erosion (25%) were the dominant braiding mechanisms (Figure
2.6). Overall, morphologic change in this run was less in total compared to previous runs,
with narrow swaths of eroding banks and dissection of mid-channel compound bars, along
with larger areas where sheets of material (presumably being eroded from proximal banks
and bars) were being deposited on the margins and heads of central bars. Tracers in
this run were seeded on the tail of a compound bar (green and blue tracers), along with
an actively eroding channel bank (orange tracers). Tracer recovery for Run 3 was 36%,
with the majority of tracers being recovered between 2 and 3.5 m downstream at locations
corresponding to the head and margins of the large mid-channel compound bar complex
(Figure 2.6). There was a small non-movement mode of blue tracers (n = 9) that were not
mobilized downstream from the seed location. While a mix of tracer colors were found on
the head of the compound bar, differentiation of deposition locations based on tracer color
was found with regard to those tracers deposited on the margins of the compound bar, with
tracers originally seeded on the river right side of the anabranch leading into the compound
bar (orange) largely remaining on the river right side of the bar, and tracers seeded on the
river left side of the anabranch leading into the compound bar (green/blue) largely being
deposited on the river left margin of the compound bar.
2.3.4 Run 4

Run 4 was net depositional, with a net sediment imbalance of +18% (Figure 2.7). Central bar development (57%), lateral bar development (14%) and bank erosion (14%) were the dominant braiding mechanisms. Morphologic change during Run 4 was highlighted by an elongate, narrow band of bank erosion on the river right margin of the braid plain, along with trimming of the margins and tail of the large mid-channel compound bar complex. Deposition of material (presumably sourced from the area of bank erosion immediately upstream) was seen in what was formerly a large chute dissecting this compound bar (Figure 2.7). In this run, tracers were seeded in close proximity in a plane-bed area of the main channel anabranch 2 m upstream of the compound bar. 33% of tracers were recovered, including a large non-movement mode (n = 27) of orange tracers which remained in their original seed location. The mobilized tracers were recovered over a relatively large downstream distance, with a peak in tracer recovery from 2-4 m downstream of the seed location corresponding to the head of a large compound bar, with some tracers being deposited at sub-diffuences of smaller anabranches dissecting this compound bar (Figure 2.7). A smaller compound bar on the river-right side of the braid plain, formed by the dissection of a larger compound bar into two surfaces by a large anabranch received 20 additional tracers deposited at 4.5 - 5 m downstream of their seed locations, with the majority of these being deposited at the head of the compound bar (Figure 2.7). Four additional tracers were recovered ∼10 m downstream from the margins of a compound bar.

2.3.5 Run 5

Run 5 was also net depositional (+13%; Figure 2.8). Dominant braiding mechanisms in Run 5 included the development of central (36%) and lateral bars (27%), along with bar sculpting (18%). Geomorphic change during this run was largely the result of the erosion of banks and bar margins, along with the chute cutoff of point bars contributing material that was either deposited directly within anabranches or which was agglomerated onto the margins of developing point bar surfaces. Tracers in Run 5 were seeded in a head-cutting
chute dissecting a point bar (orange) and on the margin of this point bar (green). We initially attempted to seed blue tracers on an actively eroding cutbank, but observed that many of these tracers were immediately washed downstream along the water surface rather than being transported as bedload, and as such we do not report the results of blue tracer recovery here as their transport distances may not reflect bedload dynamics accurately. Rather, we seeded 50 yellow tracers to the same eroding bank with 7 minutes remaining in the run in an attempt to account for the lack of blue tracers in this run, and report their recovery here. 45% of the tracers were recovered. With regard to yellow tracers, a large non-movement mode of particles was observed at the seed location (n = 22). A smaller group of yellow particles were transported downstream from the seed location and observed to saltate along the bank, where they were recovered along the river right bank of the anabranch from 1 - 1.5 m downstream of the seed location. Both orange and green particles exhibited non-movement modes, with 40 particles total being recovered at their original seed locations from this group. The remaining particles were transported 2.5 - 4 m downstream, where they were deposited on the margins of a point bar and a compound bar; of the 200 orange and green particles, only 3 were recovered from areas between the seed locations and the dominant point bar/compound bar margin deposition location (Figure 2.8).

2.3.6 Synthesis of Results

Across all five runs, the majority of tracers were recovered in areas of very thin-mantled deposition (Figure 2.9). When examining mobilized tracer recovery locations for all 351 mobilized and recovered particles overlain on our thresholded DoDs (Section 2.3), only 2% of tracers were recovered in areas that the DoDs show as being net erosional over the course of a run. 12% were found in depositional areas, while 86% of all recovered tracers were found in areas that exhibited neither detectable deposition nor erosion. Although this at first may seem counterintuitive, examining tracer deposition locations using the raw (un-thresholded) DoDs reveals that 76% of mobilized tracers were deposited in low-magnitude depositional
areas - the discrepancy between this and our original estimate of 12% of recovered tracers in depositional areas is explained by the conservative thresholding process (Equation 2; \( p = 0.01 \)) used in our DoD analyses having removed many areas of shallow erosion and deposition from the final DoD. Indeed, of all recovered tracers, the mean elevation change from the raw DoDs associated with tracer recovery locations was 0.002 m of deposition (Figure 2.9). This low-magnitude deposition associated with many recovered tracers would have been thresholded out of our DoDs via our probabilistic thresholding method (Section 2.2.3).
**Figure 2.4.** Flume Run 1 Results: Tracers, Path Lengths, and Geomorphic Change. Top image shows thresholded DoD, along with seed locations and recovery locations of all tracers. Middle image shows a zoomed-in section of tracer map; path-length distribution for all tracers shown at right. Bottom image shows braiding mechanisms along with volumetric contribution of each mechanism to total geomorphic change.
**FLUME RUN 2: RESULTS**
**DEM-OF-DIFFERENCE & TRACERS**

![Image of FLUME RUN 2 Results: Tracers, Path Lengths, and Geomorphic Change.](image)

**Elevation Change (m)**
- High: 0.02
- Low: -0.02

- Tracer Seed
- Recovered Tracer

**VOLUMETRIC EROSION:** 2750 CM$^3$
**VOLUMETRIC DEPOSITION:** 2484 CM$^3$
**VOLUMETRIC IMBALANCE:** -267 CM$^3$
**PERCENT IMBALANCE:** -3%

**BRAIDING MECHANISMS**

- Central Bar Development
- Bar Sculpting
- Channel Incision
- Chute Cutoff
- Confluence Pool Scour
- Lobe Dissection
- Overbank Sheets
- Questionable or Unresolved
- Transverse Bar Conversion
- Bank Erosion

**PATH LENGTH DISTRIBUTION**

**Figure 2.5.** Flume Run 2 Results: Tracers, Path Lengths, and Geomorphic Change. Refer to Figure 2.4 caption for details.
FLUME RUN 3: RESULTS
DEM-OF-DIFFERENCE & TRACERS

Elevation Change (m)
- High: 0.01
- Low: -0.02
- Tracer Seed
- Recovered Tracer

VOLUMETRIC EROSION: 1145 CM³
VOLUMETRIC DEPOSITION: 3247 CM³
VOLUMETRIC IMBALANCE: +2108 CM³
PERCENT IMBALANCE: +24%

PATH LENGTH DISTRIBUTION

BRAIDING MECHANISMS
Mechanism
- Central Bar Development
- Bar Sculpting
- Channel Incision
- Chute Cutoff
- Confluence Pool Scour
- Lobe Dissection
- Overbank Sheets
- Questionable or Unresolved
- Transverse Bar Conversion
- Bank Erosion

Figure 2.6. Flume Run 3 Results: Tracers, Path Lengths, and Geomorphic Change. Refer to Figure 2.4 caption for details.
Figure 2.7. Flume Run 4 Results: Tracers, Path Lengths, and Geomorphic Change. Refer to Figure 2.4 caption for details.
Figure 2.8. Flume Run 5 Results: Tracers, Path Lengths, and Geomorphic Change. Refer to Figure 2.4 caption for details.
Figure 2.9. Geomorphic Unit Associations of Recovered Tracers. Left-hand chart shows geomorphic units where tracers were deposited; center chart shows surfaces where tracers deposited on bars were found, normalized by area of surfaces across all runs. Histogram on right shows elevation changes associated with mobilized tracer recovery locations.
2.4 Analysis

This study documents the striking influence of channel bars, both bank-attached and mid-channel, in acting as deposition sites for particles in transport, and as such the path lengths of bed particles closely correspond to the locations of these discrete geomorphic units, suggesting an inherent coupling between sites of erosion and deposition during mobilizing events. Over the course of five flume runs in a continually-evolving braided channel, we found that compound bars, diagonal bars, and point bars acted as deposition sites for 81% of particles that were entrained. Further, of the particles that were deposited on bars, 70% of these were either recovered on bar heads or bar margins, with only 30% being recovered from the tops, tails, or chutes on bar surfaces. The importance of bar heads as depositional sites for particles has been noted in previous tracer work conducted in this same flume \cite{Gitto2012}. To account for areal differences between bar surface units, we normalized tracer deposition by the spatial extent of each surface on any bar that received tracer particles (Figure 2.9). Despite accounting for only 26% of total bar surface area, bar heads and margins received notably more tracers (22 tracers/m$^2$ and 27 tracers/m$^2$, respectively) than the other three surfaces. Conversely, bar tops, which comprised 53% of bar surface area, received just 3 tracers/m$^2$ (Figure 2.9).

In terms of elevation changes associated with tracer deposition, we observed that the majority of deposition sites (86% of all mobilized tracers) were areas of low-magnitude elevation change that were initially thresholded out from our DoDs as indistinguishable from survey noise. We did not find a strong influence of particle source location on subsequent depositional location classification, but did observe instances in which source location appeared to play a role in determining the path that particles took in transit downstream to their eventual deposition sites, thus influencing the particles’ path lengths.

This research points to the importance of bars in acting as trapping sites for sediment in transit downstream, with a large majority of recovered mobilized particles (80%) being deposited on compound, point, and diagonal bars. Additionally, we anecdotally note that during flume runs, injected tracers were frequently entrained immediately upon introduc-
tion to the flow and reached bar heads or margins within a minute of seeding, indicating the persistence of deposition sites throughout the course of an event. These observations highlight the utility of physical models, as this tight coupling and rapid transfer of particles between erosion and deposition sites is not often known (or knowable) from field-based studies of rivers at competent flows. The location of downstream deposition surfaces, dominantly bar heads and margins, remained relatively consistent in runs 1-3 (Figures 2.4-2.6), and as a result, the path-length distributions from these runs closely mirror one another. However, despite changes in morphologic configuration and bar locations during runs 4 and 5 as compared to earlier runs, peaks in the path-length distributions still corresponded to the location of bar heads and margins, suggesting that the probability of particle deposition is not spatially homogeneous, but rather that the location of channel bars (in particular bar heads and margins) are sites of increased depositional probability for particles in transit. These findings underscore previous research documenting preferential deposition sites for material in transit: both Braudrick et al. [1997] and Welber et al. [2013] noted that instream wood exhibited similar depositional patterns in association with bar heads.

Previous work aimed at developing scaling relationships in braided rivers hypothesized that if bars indeed act as discrete transient storage sites for sediment traveling downstream, the spacing of bars, quantified by the distance between confluence-diffuence couplets, may provide a first-order approximation of sediment travel distances [Hundey and Ashmore, 2009]. These confluence-diffuence couplets can be considered the basic unit of braided channels [Ashmore, 2013]. Given our high-resolution topographic data along with corresponding information on path lengths, we conducted a test of this hypothesis by obtaining the average spacing between corresponding confluence-diffuence couplets in main anabranches along the tracer flowpaths. Although the number of runs is small (n = 5), the relationship between bar spacing and path length is marginally significant (p = 0.06), and over all runs the mean path length of 2.5 m, as defined by the transport distance of all mobilized tracers, is indeed closely approximated by the mean confluence-diffuence spacing of 2.3 m (Table 2.2), providing an initial confirmation of the hypotheses of Hundey and Ashmore [2009].
The fact that the path length is, on average, longer than confluence-diffuence spacing may point to the importance of bar margins in acting as deposition sites, with particles frequently moving past diffuences (e.g., bar heads) and being emplaced further downstream along bar margins. When normalized by the average main channel width across all runs (0.44 m), a confluence or diffuence occurs on average once every 5.2 channel widths, consistent with the pool spacing (5-7 channel widths) that forms the fundamental geomorphic unit in many single-thread channels [Leopold and Wolman, 1957]. Because our flume runs sought to investigate the consistency of path length - morphologic unit relationships, we used a constant discharge and channel slope for all runs. As such, there was relatively little variability in the confluence-diffuence spacing. Since relatively few prior studies have documented this relationship across a range of river scales [Hunay and Ashmore, 2009], the relationship between unit spacing and particle path lengths is deserving of further research.

In our flume studies, Runs 1, 2, and 3 generated path-length distributions that contained downstream peaks corresponding to the location of compound bar heads and margins, while containing small or nonexistent non-movement modes and short-distance transport modes. These path-length distributions could be described using single-peaked Cauchy functions as in Pyrce and Ashmore [2003a]. Conversely, Runs 4 and 5 contained large peaks in the path-length distributions corresponding to an immobile fraction of material, while also containing downstream peaks that could be related to compound and point bars. These path-length distributions could be approximated using bi-modal functions that combine aspects of gamma and Cauchy distributions [Pyrce and Ashmore, 2003a]. These fitted functions are shown in Figures 2.4-2.8 and assessed using the test statistic (D) via Komolgorov-Smirnov goodness-of-fit tests. In runs 2 and 4, smaller downstream peaks are visible in path length distributions that may correspond to subsequent particle deposition on downstream bars [Pyrce and Ashmore, 2003a].

The delineation of bar surfaces (heads, margins, tops, tails, chutes) revealed the importance of bar heads and margins in acting as deposition sites for tracer particles. 70% of all particles that were mobilized and recovered were found on either the heads of bars
Table 2.2. Relationship Between Confluence-Diffuence Spacing and Path Lengths

<table>
<thead>
<tr>
<th>Run Number</th>
<th>Avg. Con.-Diff. Spacing (m)</th>
<th>Avg. Path Length (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Blue: 1.7</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Green: 2.0</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Orange: 2.5</td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>2.4</td>
<td>All Tracers: 2.2</td>
</tr>
<tr>
<td></td>
<td>Blue: 2.2</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Green: 2.2</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Orange: 2.2</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>2.4</td>
<td>All Tracers: 2.2</td>
</tr>
<tr>
<td></td>
<td>Blue: 2.3</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Green: 3.1</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Orange: 2.9</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>2.2</td>
<td>All Tracers: 2.8</td>
</tr>
<tr>
<td></td>
<td>Blue: 4.9</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Green: 5.0</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Orange: 2.4</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>2.6</td>
<td>All Tracers: 3.3</td>
</tr>
<tr>
<td></td>
<td>Green: 2.7</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Orange: 2.5</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Yellow: 1.0</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>2.0</td>
<td>All Tracers: 2.2</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Average of All</strong></td>
<td><strong>2.3</strong></td>
<td><strong>2.5</strong></td>
</tr>
</tbody>
</table>

(29%) or the margins of bars (41%). An additional 18% and 9% of particles were recovered from tops and chutes of bars respectively, whereas just 3% of particles were recovered from bar tails. It is likely that these results are at least in part a reflection of the relatively low-magnitude flows that were used in our flume runs. The flows of 2 ls\(^{-1}\) corresponded to the formative discharge of this braided channel planform, but did not completely inundate the braidplain over the duration of the run. Care was taken at the start of each run to delay the introduction of tracers until the flow was generally confined to pre-existing channels, rather than moving through the system primarily as sheet wash over the braidplain. The confinement of flow to the channels likely contributed to preferential deposition of particles
on bar heads or margins, which were either inundated or proximal to flow, while also resulting in less common deposition of particles on bar tops or chutes via transport of tracer particles onto and over bar surfaces. The lack of particle deposition on bar tails is also of interest, especially when compared to the frequency with which tracers were recovered from bar margins and bar heads. This likely indicates that the rapid decrease in transport competence via either a reduction in flow depth (in the case of bar heads) or flow velocity (in the case of inner-bend bar margins) exerts a pervasive influence on particle transport such that few particles are able to bypass these likely deposition sites and be mobilized downstream to bar tails. Finally, the tracer size in these runs (tracer $D_{50} = 2.4$ mm as compared to bed $D_{50} = 1.2$ mm) may have prevented these coarser particles from being transported to and deposited on bar tails.

We did not observe any influence of tracer seed location (i.e. geomorphic unit) on the units in which tracers were deposited. In Runs 1, 2, and 3 the path-length distributions of individual tracer colors largely mirrors the overall path-length distribution for all recovered tracers. Rather, any differentiation in deposition locations of tracers appeared to be the result of hydraulic influences that affected which anabranches tracer particles were mobilized through. This effect is visible in runs 3, 4, and 5, where tracers were recovered in groups often differentiated by the color of the tracer. In runs 3 and 5 the paths taken by orange/green tracers are markedly different from those taken by blue tracers (and yellow tracers, in the case of run 5). Visual analysis of tracer paths from videography captured during the runs, but not presented here, confirmed the differentiation of paths taken by tracers as a result of their initial seed location. Nevertheless, no matter the seed location of the tracers, the geomorphic units acting as particle deposition locations did not vary - for all tracers, compound, point, and diagonal bars were the dominant deposition site, whereas erosional/transfer units such as anabranches and confluence pools received comparatively few tracer particles. Additionally, the relative mobility of the particular patch at which tracers were seeded did not appear to play a role in driving variations in particle mobility; large numbers of immobile patches of tracers were seen in runs 4 and 5, for example. How-
ever, once these tracers became entrained in the flow and deposited downstream, they were found on bar surfaces alongside counterparts sourced from highly mobile seed locations.

2.5 Discussion

This research could influence the way we collect data on braided rivers and improve our fundamental understanding and conceptualization of sediment transport processes occurring therein. This study is one of the first documented uses of SfM to generate elevation models and detect geomorphic change in a physical model setting. The use of SfM has dramatically increased in recent years as a low-cost and rapid method for surveying topography and monitoring geomorphic change in the subaerial portions of the fluvial corridor [Westoby et al., 2012; Fonstad et al., 2013]. More recently, Javernick et al. [2014] have explored leveraging orthorectified imagery created via SfM in combination with spectral-depth correlations to extract bathymetry and obtain a complete DEM of the fluvial environment. While the images in this study were captured in a controlled setting where water could be drained prior to photography, the rapidity and ease of generating DEMs with very high spatial resolution (0.003 m) and the reliability with which we were able to detect morphologic changes (at Δz levels corresponding to tracer D50) point to the potential utility of this technique in other fluvial physical experiments.

Additionally, this study provides a valuable dataset describing particle path lengths, an area in which others have noted a lack of reliable data at the flood event scale [Pyerce and Ashmore, 2003a]. Although our understanding of the processes driving the development of anabranches and bars in braided rivers has existed for some time [e.g., Leopold and Wolman, 1957; Ashmore, 1991], our conceptual understanding of the grain-scale nature of transport processes and their relationship and associated feedbacks with channel morphology remain underdeveloped in these systems. Previous research documenting morphologic controls on particle travel distances [Pyerce and Ashmore 2003a, 2003b] has hypothesized that the bar spacing in braided channels (specifically the distance between confluence/diffluence couplets) may be a predictor of sediment path lengths. If bars act as discrete transient
storage sites for particles undergoing episodic downstream transport, the mean distance between those bars should be closely related to the average path length of sediment. This research provides an important first step in confirming those initial hypotheses, finding that characteristic confluence/diffluence spacing closely approximates average particle path length for all flume runs (Table 2.2).

Although this research did not directly explore the effect of flood magnitude or duration on particle path length, the simulated flows potentially offer insight with regard to the persistence of deposition sites (and hence consistency of particle path lengths) throughout an event of intermediate magnitude. During each flume run, we noted that tracer particles mobilized downstream from seed locations were quickly (generally within ~1 minute of seeding) deposited on bar heads and margins and remained there throughout the course of the run. As such, we hypothesize that particle path length distributions may be independent of high flow magnitude or duration, insofar as that high flow does not result in large-scale reworking of bar surfaces where particles are stored; thus, for particularly high-magnitude or long-duration events that rework bars, path length distributions may be significantly altered through the course of the flood. Additionally, characteristic path lengths may be altered with varying discharge even in the absence of morphodynamic evolution, as new networks of anabranches with variable confluence-diffluence spacing are activated at increasing stage. Conceptual models detailing the effect of increasing flood magnitude on resultant path length distributions have been put forward by previous researchers [Pyrce and Ashmore, 2003a], and given the relative stability of these distributions in our short-duration, intermediate-magnitude experiments, we believe this question of path length stability is deserving of further study, potentially via high-speed videography and particle tracking.

The relationship between channel morphology and particle path lengths found in this research may help to inform the underlying conceptual framework with regard to numerical modeling of braided systems. While it is true that morphodynamic models exist that can predict the evolution of braided streams and even the transport of individual particles, these frequently require large computational overhead for computation of the associated fluid dy-
namics [Nicholas, 2013], or drastically simplify this at the expense of realism with respect to
topographic reconstruction [Murray and Paola, 1994; Thomas and Nicholas, 2002; Nicholas,
2005]. We believe that the conceptual understanding of likely depositional sites for tracers,
along with the relatively small influence of seed location on recovery location developed in
this study points to a fundamental form-process linkage in braided rivers and paves the way
for computationally-efficient morphodynamic models which drive sediment transport and
channel evolution as a function of particle path lengths. Using an a priori understanding
of channel geomorphic units (e.g., types of bars, banks, and anabranches) encountered by
entrained sediment in transport, probabilistic path length distributions may be derived that
apply the findings of this work, thereby increasing deposition probability in certain loca-
tions (bar heads and margins, for example) while predicting a lower deposition probability
in other units (e.g., anabranches, confluence pools, bar tops/tails). The development of such
models will undoubtedly require the collection of more event-scale datasets such as the one
described here by which we can more fully develop a conceptual framework for downstream
sediment transport in braided channels.

Perhaps the chief limitation of our study is the methodological inability to recover
(a) deeply buried or (b) exported particles and include their transport distances in our
path-length calculations. Nevertheless, the agreement between our research and previous
work that sought to recover buried particles [e.g., Pyrce and Ashmore 2003a, 2003b] may
indicate that tracers recovered from the channel bed are quite representative of their more
deeply-buried counterparts. It is also possible that tracers remaining from previous tests
could have been exhumed and recovered in subsequent runs, but we attempted to minimize
this effect by thoroughly searching the flume bed between runs with the aid of an ultraviolet
lamp. Nevertheless, these difficulties often arise in visually-based tracer recovery studies in
field settings as well, and our recovery rates are similar to those computed in field studies
[Ashworth and Ferguson, 1989; Schmidt and Ergenzinger, 1992; Hattingh and Illenberger,
1995; Sear, 1996]. As such, future work that employs alternative techniques for tracer re-
cover (e.g., magnetic particles, radio-tagging; [Schmidt and Ergenzinger, 1992; Habersack,
2001; Bradley and Tucker, 2013]) may increase tracer recovery rates and help to account for particle burial or export and thus modify our understanding of path-length channel form relationships.

2.6 Conclusions

This study employed high-resolution digital elevation models derived from structure-from-motion photogrammetry, novel techniques for geomorphic change detection and mechanistic budget segregation, and visually-tagged sediment tracers to document the coupling between particle path lengths and channel morphology in a physical model of a braided river. In so doing, we sought to conduct one of the first studies that explicitly quantifies both morphodynamic channel change and particle transport and draw linkages between them during competent flows.

The results of this study point to the influence of bars in acting as deposition sites for particles in transport, with over 80% of tracers that were recovered being found in association with compound, lateral, diagonal, and point bars. Of these tracers, 70% were recovered from bar heads and bar margins, with the remainder being recovered from bar tops and chutes; just 3% of recovered tracers were found on bar tails. Neither seed location nor relative mobility of a seed patch appeared to play a large role in determining the downstream deposition sites of tracers. Once particles were in transport, they were often found intermingled in similar geomorphic units with particles sourced from other tracer patches. Instead, seed location appeared to play a role in influencing the anabranches through which particles were transported on their way downstream, thereby indirectly influencing deposition sites, although geomorphic units remained similar despite differences in the area in which particles were deposited. Most recovered particles were deposited in areas of thin deposition (mean $\Delta z = 0.002$ m, $n = 351$) that were initially thresholded out of DoDs as being indistinguishable from survey noise.

The results of this study point to the value in using novel techniques for generation of digital elevation models, particularly the structure-from-motion approach employed here.
Additionally, this research furthers our conceptual understanding of form-process linkages between channel morphology and particle path lengths in braided streams, which has the potential to improve the efficiency of morphodynamic models for braided rivers through the use of simplified relationships between channel morphology and sediment travel distances. Finally, despite our inability to recover tracers which were deeply buried or exported (as a result of our post-facto visual tracer recovery from photographs), we argue that the signal in our data trumps the potential noise and suggests a clear morphologic influence on particle path lengths that is deserving of further exploration, particularly using tracer recovery techniques that have the potential to recover buried particles.

2.7 Acknowledgments

Alan Kasprak, Joseph Wheaton, and James Hensleigh were funded by a National Science Foundation grant (7086465). Peter Ashmore’s research program and flume lab are supported by an NSERC Discovery grant. Construction of the laboratory flume was funded by Canada Foundation for Innovation and Newalta Resources. Sarah Peirce is funded by the Vanier Canada Graduate Scholarship. Philip Bailey (North Arrow Research) and Sara Bangen (Utah State University) provided guidance with regard to DoD generation. Yannick Rousseau (University of Western Ontario) provided logistical support during flume work, and Alex Gitto (University of Western Ontario) provided methodological groundwork for our experiments. We thank Nate Hough-Snee, Martha Jensen, Eric Wall, Gary O’Brien, and Konrad Hafen (Utah State University) for critical and valuable commentary on this research. Alexander Densmore, Walter Bertoldi, and Murray Hicks provided constructive reviews that greatly improved this manuscript. Data used in this research can be accessed at http://dx.doi.org/10.6084/m9.figshare.1288618.
References


CHAPTER 3

COMING TO GRIPS WITH MODEL IMPERFECTION: MORPHODYNAMIC MODELS AS EXPLORATORY TOOLS FOR UNDERSTANDING BRAIDED RIVER DYNAMICS

Abstract

Numerical models that predict channel evolution through time are an essential tool for investigating processes that occur over timescales which render field observation intractable. However, available morphodynamic models generally take one of two approaches to the complex problem of computing morphodynamics, resulting in drastic oversimplification of the relevant physics (e.g. cellular models) or faithful, yet computationally intensive, representations of the hydraulic and sediment transport processes at play. The practical implication of these approaches is that river scientists must often choose between unrealistic results, in the case of the former, or computational demands that render modeling realistic spatiotemporal scales of channel evolution impossible, in the case of the latter. Here we present a new modeling framework that operates at the timescale of individual competent flows (e.g. floods), and uses a highly-simplified sediment transport routine that moves volumes of material according to morphologically-derived characteristic transport distances, or path lengths. Using this framework, we have constructed an open-source morphodynamic model, termed MoRPHED, which is here applied, and its validity investigated, at timescales ranging from a single event to a decade on two braided rivers in the UK and New Zealand. We do not purport that MoRPHED is the best, nor a perfect, tool for modeling braided river dynamics at this range of timescales. Rather, our goal in this research is to explore the utility, feasibility, and sensitivity of an event-scale, path-length-based modeling framework for predicting braided river dynamics. To that end, we further explore (a) which processes are naturally emergent and which must be explicitly parameterized in the model, (b) the sensitivity of the model to the choice of particle travel distance, and (c) whether an

\footnote{This chapter is in preparation for submission to *Earth Surface Processes and Landforms* and is co-authored with Konrad Hafen and Joseph Wheaton (Utah State University) and James Brasington (Queen Mary, University of London).}
event-scale model timestep is adequate for producing braided channel dynamics. The results of this research may inform techniques for future morphodynamic modeling that seeks to maximize computational resources while modeling fluvial dynamics at the timescales of channel change.

3.1 Introduction

Numerous processes in riverine environments occur over time and space scales that render traditional field-based observation impractical or impossible (Gurnell et al., 2009). These fluvial dynamics include channel migration (Hooke, 1995; Black et al., 2010), shifts in channel form (Kondolf et al., 2002), and alterations in hydrology and/or sediment delivery (Grams and Schmidt, 2005). All of these dynamics frequently occur on timescales ranging from years to centuries, and channel response to hydrologic and sediment regime shifts may manifest across a variety of spatial scales ranging from individual channel units (e.g. meters) to reaches spanning several kilometers. In such instances where the spatiotemporal scale of channel response renders field-based methods of observation intractable, representation of the fluvial environment using numerical models is invaluable both in terms of disentangling the relative efficacy of competing processes acting to shape channels and predicting future channel response to geomorphic forcings (Nicholas and Quine, 2007; Gurnell et al., 2009).

Despite the immense value of numerical models in explanation and prediction of fluvial processes, the timescales at which channel evolution occurs render most available morphodynamic models (those which predict changes in channel form over time) impractical. Historically, one way of dealing with this problem has been to simplify the physics involved in modeling, giving rise to the so-called 'reduced complexity' or 'cellular automata' models (RC/CA; Murray and Paola, 1994; Coulthard et al., 2002; Thomas and Nicholas, 2002). These models simplify the transport of water and sediment across a cellular network representing the riverscape using a rule set governing each process involved. Because these rule sets are simplified representations of the physics involved in hydraulics and sediment transport, RC models achieve a great deal of computational efficiency, allowing calculations
over large spatiotemporal extents (e.g. kilometer-scale, decadal-to-centennial timescales; Nicholas and Quine, 2007; Thomas et al., 2007). Yet this computational efficiency comes at the expense of field realism; because of the simplified nature of the physical processes, particularly the inability to conserve hydraulic momentum leading to inaccurate representation of pool dynamics and meander migration (Nicholas and Quine, 2007), reduced complexity models often fail to reproduce observed channel behavior at the spatial scales of change.

On the other hand, the subset of morphodynamic models driven by computational fluid dynamics (CFD; Bates et al., 2005) involve hydrodynamic components that approximate the solution of the Navier-Stokes Equations and subsequently drive sediment transport. To ensure computational stability, morphodynamics are typically computed by solving a form of the equation of sediment continuity (Exner Equation; Paola and Voller, 2005) at fine time steps (seconds-minutes) using solutions from the equations of motion, or Navier-Stokes equations presented below in three dimensions (Equations 3.1-3.3), which relate changes in momentum (left-hand terms) in time and space to the cumulative surface and body forces acting on the fluid (right-hand terms).

\[
\frac{\partial (\rho U_x)}{\partial t} + \sum U_j \frac{\partial (\rho U_x)}{\partial x_j} = \rho F_{\text{vol}x} - \frac{\partial P}{\partial x} + F_{\text{visc}x}
\] (3.1)

\[
\frac{\partial (\rho U_y)}{\partial t} + \sum U_j \frac{\partial (\rho U_y)}{\partial x_j} = \rho F_{\text{vol}y} - \frac{\partial P}{\partial y} + F_{\text{visc}y}
\] (3.2)

\[
\frac{\partial (\rho U_z)}{\partial t} + \sum U_j \frac{\partial (\rho U_z)}{\partial x_j} = \rho F_{\text{vol}z} - \frac{\partial P}{\partial z} + F_{\text{visc}z}
\] (3.3)

In the generalized form of the Exner Equation below (Equation 3.4), bed elevation \((z)\) through time \((t)\) is a function of the sediment supplied from upstream \((V_s)\), the divergence of the sediment flux through the reach boundaries \((Q_s)\), and the porosity of the deposited sediment \((\gamma_p)\). This reliance on rapid calculation of morphodynamic evolution comes at the cost of vastly increased computational overhead, making CFD-driven morphodynamic models suitable only over fine spatiotemporal scales for most users (e.g. hours-months at
meter-scale resolution; Coulthard and Van De Wiel, 2012; Ferguson, 2007).

\[
\frac{\partial z}{\partial t} = \frac{1}{1 - \gamma_p} \left( \frac{\partial V_z}{\partial t} + \nabla \cdot Q_s \right)
\]  

(3.4)

As a result of the shortcomings inherent to both RC and CFD models, there exists a knowledge gap whereby there are a lack of morphodynamic models optimized for use at 'intermediate' spatiotemporal scales. Practically speaking, these scales translate to meter-scale resolution over kilometer-scale reaches (i.e. resolving the 'mesoscale'; Crowder and Diplas, 2000; Newson and Newson, 2000), with the ability to model channel morphodynamics over years to decades. Modeling at these intermediate scales is particularly important, as these are the typical timescales of channel adjustment in response to hydrologic and sediment supply forcings (e.g. Montgomery and Buffington, 1998; Gurnell et al., 2009), and the spatial scale at which ecologic response to channel form is commonly observed and related to physical habitat metrics (Minns et al., 1996; Rosenfeld, 2003).

Despite the long-term predictive ability of RC/CA models (Nicholas and Quine, 2007; Thomas et al., 2007), the current state-of-the-science in morphodynamic modeling continues to consist of a deterministic, reductionist approach that is underlain by the continuity equation (3.4) and relies on advances in computational power to drive understanding of morphodynamic processes forward. We do not argue that dependence on sediment continuity yields unreliable or invalid results, as CFD-based models have been used reliably in coastal and estuarine settings for decades (De Vriend et al., 1993; Bates et al., 2005), with a more recent yet robust application to fluvial systems (Mosselman et al., 2000; Rinaldi et al., 2008). More recent fluvial applications of CFD-based morphodynamics have shown remarkable promise in their ability to model the continuum of single-thread to braided channel planforms over centennial timescales (Nicholas, 2013), although these model domains require the use of supercomputing resources. We do believe that short of remarkable advances in computational power for individual users, the reductionist and purely dynamics-based approach taken with CFD-driven models is not a tractable way forward in modeling fluvial morphodynamics at intermediate spatiotemporal scales.
Contrary to previous modeling efforts, we argue that neither CFD nor RC models need to exist in a vacuum. Instead, we believe that the fusion of CFD and RC-based morphodynamic modeling may present a novel way forward, in that the high spatial fidelity afforded by CFD-driven models may be coupled with a simplified, empirically-derived rule set for sediment transport and morphodynamic channel evolution (Nicholas and Quine, 2007). In predicting channel evolution via a combination of dynamics- and kinematics-based approaches, the result of this fusion has the potential to produce models capable of capturing the 'intermediate' spatiotemporal scales that currently elude available morphodynamic approaches. As such, this paper presents a new, hybrid morphodynamic model termed the Model of Riverine Physical Habitat and Ecogeomorphic Dynamics (MoRPHED) that employs two-dimensional CFD hydraulics in concert with an event-based rule set for sediment transport that leverages previous research on particle path lengths, or the characteristic distance traveled by sediment particles during a flood (Pyrce and Ashmore, 2003a,b; Kasprak et al., 2015), to vastly simplify sediment transport. Previous research on particle path length distributions has largely been conducted in gravel-bed braided rivers, and as such MoRPHED is developed for these systems; however, the existence of characteristic particle travel distances has also been documented in meandering single-thread channels (Pyrce and Ashmore, 2003b), meaning it may be possible to apply the underlying theory to a wide variety of channel planforms.

We do not purport that MoRPHED is the best available morphodynamic model, nor do we attempt to argue that its outputs faithfully reproduce observed field behavior in all instances. Herein, we test the model using high-resolution topographic datasets at the reach scale over a variety of timescales, from single floods to a decade. While we attempt to demonstrate the utility and validity of the model at a variety of space and time scales, our goal is not to present this software as the endpoint of RC-CFD fusion modeling; far from it, we view this model as a potentially important first step in this approach and have designed the software as a modular and exploratory framework for testing the way we represent braided river morphodynamics algorithmically. The shift to models such as
the one presented herein is, we contend, a novel approach, and the optimal methods for representing the kinematics of traditionally continuum-based processes at coarse temporal scales (e.g. bank erosion and bed scour) are as yet unknown. As such, the degree to which we are able to represent emergent properties of the fluvial environment, along with the model’s sensitivity to contrasting process representations and current shortcomings with regard to its validity, are explored as part of the multi-scalar validation presented here.

3.2 The Model

As with previously-developed morphodynamic models, the model used here (MoRPHED v. 1.1; github.com/morphed) contains routines for simulating hydraulics and uses these calculations to drive sediment transport. This section details the methods used in each of these components, along with ancillary routines such as the parameterization of model boundaries, sediment grain size, and bank erosion. A flowchart of model operation along with required/optional inputs and outputs are shown in Figure 3.1, and these components are discussed throughout this section.

3.2.1 Hydraulics

The model’s hydraulic component is driven using the freely-available, open-source Delft3D software (Version 4.00.01, Deltas, Delft, Netherlands). Delft3D solves the shallow-water form of the Navier-Stokes equations, and herein we employed the model in two-dimensional (depth-averaged) form, as this provided an ideal compromise between computational efficiency and the ability to resolve hydraulics, specifically flow depth, direction, and bed shear, at the cellular scale of our DEMs (Lane et al., 1999). Although three-dimensional simulations using Delft3D are possible, the parameterization, validation, and computational overhead associated with three-dimensional modeling precludes their use in this model (Lane et al., 1999; Brasington and Richards, 2007). For all modeling, we employed Cartesian orthogonal grids generated using the RGFGRID module of the Delft3D suite and kept constant throughout a modeled event series, and adjusted the model time
step to satisfy the Courant-Friedrichs-Levy condition and ensure computational stability. Models were run at a steady upstream discharge and were allowed to run to steady state (no observed change in depth, velocity, or inundation extent in QUICKPLOT model postprocessor), so as to simulate the hydraulics of a given flood without needlessly extending computational time.

For all simulations, discharge was specified at the upstream boundary and a corresponding water surface elevation was set at the downstream boundary. Horizontal eddy viscosity ($\nu$) was set at 0.1 s/m$^2$ (Williams et al., 2013). We set a constant roughness value
based on the Colebrook-White equation to determine the 2D Chezy coefficient:

\[
C_{2D} = 18 \log_{10} \frac{12H}{k_s}
\]  

(3.5)

where \(H\) is water depth and \(k_s\) is the Nikuradse roughness length, which can be described in terms of a factor \((\alpha_x)\) of the characteristic grain diameter as:

\[
k_s = \alpha_x D_{84}
\]  

(3.6)

Here, we used \(D_{84}\) as the characteristic grain size as it provides an estimate of coarse grain influence on the flow field. Using pre-surveyed grain size distributions for both of the modeling sites used in this paper (see Section 3.4), the Rivers Rees and Feshie (Hodge et al., 2009; Williams et al., 2013), we computed \(k_s\) using (3.6) and \(\alpha\) of 2.9, as averaged from a range of gravel-bed rivers studied in Garcia (2006) as 0.1 (Rees) and 0.29 (Feshie); the larger \(k_s\) on the Feshie is primarily the result of the larger bed material grain size in that reach as compared to the Rees.

Delft3D requires an input DEM, along with simulated water discharge and boundary conditions. Downstream water surface elevation for each modeled discharge was used to parameterize the hydraulic model boundary and was calculated by determining reach-scale conveyance associated with reach-average slope and roughness. Although numerous hydraulic variables can be computed and exported from Delft3D, here we used (a) water depth, (b) flow velocity resolved into streamwise and lateral components, and (c) bed shear stress.

MoRPHED is an event-scale model, predicting channel evolution at the scale of individual floods. We do this for two reasons, (a) because the calculation of morphodynamics at coarser intervals allows for greatly reduced computational overhead associated with the model, and (b) because we argue that modeling at finer intervals, while allowing the ability to capture rapid transient events such as prograding bedload sheets and bank retreat during the course of a single flood, is difficult if not impossible to validate since the most common
data geomorphologists have describe channel form before and after a single event (Bertoldi et al., 2010; Williams et al., 2011; Mueller et al., 2014), along with sediment transport distances resulting from that event (Pyrce and Ashmore, 2003a,b; Snyder et al., 2009; Kasprak et al., 2015).

Hydraulics were computed as a single constant discharge corresponding to the peak of a flood. The exception to this is a case study contained within this research (Section 3.6.2) wherein we discretized the hydrograph of a single flood event into three discrete steady discharges corresponding to the rising limb, peak, and falling limb. This discretization of the hydrograph for long events was attempted so as to capture distinct stage-specific features of channels such as the dissection of bar-top chutes at falling stages (e.g. Wheaton et al., 2013), and because sediment transport capacity may change throughout a hydrograph as supply is exhausted (e.g. hysteresis in the sediment rating curve; Topping et al., 2007).

3.2.2 Bed Sediment Erosion

The model employs a critical nondimensional value of the bed shear stress (Shields stress) to determine whether sediment can be entrained at a particular location. As our model is developed and used on gravel-bed rivers, the theory and threshold values of Shields stress for entrainment have been well studied in these settings. Incipient motion for gravel occurs when the Shields stress ($\tau_*$) exceeds 0.03-0.07 (Buffington and Montgomery, 1997; Snyder et al., 2009):

$$\tau_* = \frac{\tau_B}{(\rho_s - \rho)gD}$$

where $\rho_s$ is sediment density (2650 kg/m$^3$), $g$ is acceleration due to gravity, and $D$ is the median particle size. The bed shear stress ($\tau_B$) was computed using the output from Delft3D, and the critical Shields stress for sediment mobility was set at 0.05. Because bed shear varies by large degrees over small spatial regions (Wilcock et al., 2009), which could drive large cell-to-cell variability in elevation change and resultant hydraulic instability, the model averages bed shear along flowlines delineated from Delft3D-output velocity vectors for 10 cells upstream and downstream of the cell in question. Although lateral averaging
of shear stress may additionally increase model stability, this was not explored here. The result is a single average value of bed shear stress that is used for computation of scour depth.

While most morphodynamic models compute bed elevation change using some form of the Exner Equation (Equation 3.4) of sediment continuity (Paola and Voller, 2005), which relates elevation change ($\partial z$) to the downstream divergence of sediment flux ($\nabla Q_s$) while accounting for bed sediment porosity, the model cannot employ such a method given the long time steps at which the model operates. Practically speaking, the result of using Equation 3.4 would mean extremely large depressions or mounds would be formed on the bed topography. As a potential alternative, (Montgomery et al., 1996) proposed an event-scale model to predict event-scale sediment scour depth ($D_s$):

$$D_s = \frac{Q_b}{u_b \rho_s (1 - \gamma)}$$ (3.8)

where $Q_b$ is the average bedload transport rate during the event, $u_b$ is the bedload velocity, and $\gamma$ is the bed sediment porosity. Estimating $Q_b$, while straightforward, is often inaccurate as it is a strongly nonlinear process; however, most transport relations take a power-law form (e.g. Meyer-Peter Müller equation):

$$Q_b = (\tau_b - \tau_{BC})^{1.5}$$ (3.9)

and $u_b$, the bedload velocity, can be estimated as

$$u_b = a (u_* - U_{*C})$$ (3.10)

where $u_*$, the shear velocity, is computed as

$$u_* = \sqrt{\frac{\tau_b}{\rho}}$$ (3.11)

which can be estimated directly from the bed shear stress obtained from Delft3D. The
constant \( a \) in Equation 3.10 has been studied by many researchers (Garcia, 2006), but is generally around 9. Although the accuracy of using Equation 3.8 to predict event-scale scour depth has not been explored empirically, it represents one of the only methods for predicting the depth of scour over extended timescales, and is used here in a largely exploratory manner.

3.2.3 Bank Sediment Erosion

Readily-erodible banks are a hallmark of braided rivers (Wheaton et al., 2013), leading to channel widening, the deposition of mid-channel bars, and ultimately creating a braided network of channels (Ashmore, 1991). Unfortunately, bank erosion is also one of the most difficult geomorphic processes to represent numerically (Darby and Thorne, 1996; Simon et al., 2000; Rinaldi and Darby, 2007), and most models rely on fine timesteps and/or small-scale force balances to predict bank stability. As MoRPHED is a simplified event-scale model, here we estimate lateral retreat distance by empirically scaling near-bank shear stress and bank slope to the distance of lateral retreat during a model run.

To begin, the model calculates the slope of all cells in the model domain and selects those which exceed a user-defined slope criterion, which was empirically calibrated to 7% for all simulations presented herein (Figure 3.2.1). Whether bank erosion occurs by mass failure or lateral channel migration, eroding banks are often marked by steep slopes, and as such this criteria was used as the first metric for computing bank sediment erosion. This slope delineation produces groups of cells, which are removed from the selection if the group’s area falls below a user-specified threshold (Figure 3.2.2). Here, we set this threshold area to 30 cells, again adjusting this value to mirror the size of field-observed bank erosion patches; we also observed that very small area thresholds would create discontinuous patches of lateral retreat, leading to model instability and hydraulic artifacts in subsequent runs. The bed shear stress in the surrounding cells is then sampled using a 3 x 5 neighborhood window and each cell within this neighborhood that exceeds the critical shear value for the grain size in that cell is noted (Figure 3.2.3). The bed shear stress values in these cells are averaged,
producing a single shear stress value for the delineated group of cells. Those cell groups with average shear stress below a user-defined criteria, here set to 50 N/m², are excluded. For each remaining group of cells, the model removes material from a number of cells specified by Equation 3.12, which computes the lateral extent of bank erosion \( n \), rounded to the nearest whole cell, as a function of slope \( S \) and near-bank shear stress \( \tau \); Figure 3.2.4:

\[
n = \text{round}
\left((\frac{\tau}{3} + 1) \times \frac{S}{15}\right)
\]  

(3.12)

Material is removed from cells working in a direction opposite the bank cells’ aspect (Figure 3.2.4). Eroded cells are reduced in elevation to a level equal to that of the lowest cell in the initial neighborhood surrounding the bank cell (\textit{sensu} Nicholas, 2013), and eroded sediment is immediately transported downstream and deposited in a manner identical to that described for bed transport and deposition (following section; 3.3.4).
Figure 3.2. Lateral Migration Algorithm Schematic
3.2.4 Bed/Bank Sediment Transport and Deposition

Once entrained, bed or bank sediment is mobilized downstream along flowlines which are delineated using velocity vectors from Delft3D. Although the model is inherently a cell-based (e.g. raster) model, velocity vectors need not pass through the center of each cell. Instead, the absolute coordinate of each velocity vector is calculated at a downstream interval equal to the computational domain cell size. In the field, deposition of sediment consistently occurs in diffuse patterns, such as ‘tear-drop’ forms of lobate bars, prograding bedload sheets, and thinly-mantled overbank deposits (Ashmore, 1982; Ferguson and Werritty, 1983; Wheaton et al., 2013). To mirror this diffuse deposition, the model distributes sediment within a 5 x 5 window of cells surrounding the candidate deposition cell, with the candidate cell receiving 1/3 of the deposited sediment, the adjacent eight cells receiving 1/3 of the deposited sediment, and the outer ring of 16 cells receiving the final 1/3 of deposited sediment. In the case of dry cells that occur in the 5 x 5 neighborhood, the dry cell(s)’ sediment is divided among the population of wetted cells in the neighborhood.

At each cell along the flowpath, the volume of sediment to deposit in the center cell is given by a path length distribution (Figure 3.3). In the simplest sense, this distribution details the proportion of all eroded sediment which is deposited at a particular distance downstream. These distributions have been studied by numerous researchers and found to take several forms in braided rivers. Exponential decay, or heavy-tailed distributions are marked by a large number of particles mobilized short distances downstream, and may result from floods that do not generate sufficient shear stress for particle transport across the braidplain (Pyrc and Ashmore, 2003b). During floods which are competent across large areas of the braidplain, typical path length distributions exhibit peaks which correspond to the location of likely depositional sites downstream. Kasprak et al. (2015) and Pyrc and Ashmore (2003a,b) both noted that these depositional sites were most frequently the location of bar heads (e.g. flow diffuences), and as such particle path length distributions could be readily constructed using morphometric indices which described the characteristic confluence-diffuence spacing in braided channels. MoRPHED deposits a proportion of the
Figure 3.3. Modeled Path Length Distributions. Path length distributions used in MoRPHED modeling on the River Feshie (A) and River Rees (B). Peak of the distributions corresponds to average along-flow spacing between confluence and diffuence pairs.

path-length specified volume of sediment in each wetted cell of the 5 x 5 neighborhood described above. Path-length distributions in the model can take typical field-measured forms (Gaussian and Exponential Decay; Pyrce and Ashmore, 2003b), or can take any user-specified form (e.g. multi-peaked), as specified by an input text file.

3.2.5 Sediment Import and Export

For each simulated event, the model tracks the volume of sediment passing the downstream or lateral reach boundaries. In effect, export of sediment occurs when the length of the user-specified path length distribution exceeds the downstream or lateral boundaries of the model domain (i.e. sediment is to be deposited at a coordinate outside the model grid). When this occurs, the remaining volume of sediment (the amount of eroded sediment not yet deposited along the flowline) is recorded as having been exported from the reach. Sediment import is user-specified and can be (a) set equal to the volume of sediment export during the preceding event (e.g. sediment equilibrium; Grams and Schmidt, 2005), (b) specified as a percent of sediment export during the preceding event, or (c) specified via a text file detailing volumetric sediment import during each event (e.g. sedigraph timeseries). Algorithmically, the model computes flowpaths from each wetted cell at the upstream reach boundary and distributes the total volume of imported sediment to each
cell of each flowpath as specified in the user-input path length distribution.

3.2.6 Grain Size and Stratigraphy

The model dynamically tracks the median particle size ($D_{50}$) in each model cell for each simulated event. The routine used for grain size averaging is similar to that employed by Viparelli et al. (2010). The model domain is initially divided into a series of layers (Figure 3.4) extending upward from the lowest point in the initial DEM. The active, or surface, layer may vary from cell to cell with variations in elevation, and is tracked from event to event. All layers have a constant, user-defined thickness ($Z_{LYR}$) except the active layer, which may take any thickness ($Z_{ACT}$) such that:

$$0 < Z_{ACT} < Z_{LYR}$$  \hspace{1cm} (3.13)

With regard to tracking changes in $D_{50}$ at the event scale, the simplest case is that of erosion at a cell. The new grain size in that cell ($D_{NEW}$) is simply the grain size given by the corresponding layer at the cell’s new elevation (following scour):

$$D_{NEW} = D_{LYR}$$  \hspace{1cm} (3.14)

The sediment size of the scoured material ($D_{ERODED}$) is computed as either the grain size of active layer at the eroded cell (provided scour did not extend through the active layer) or as a weighted average of the eroded material (if the scour did extend through the active layer and into subsurface layers):

$$D_{ERODED} = \begin{cases} 
D_{ACT}, & \text{if } Z_{ERODED} \leq Z_{ACT} \\
\frac{\sum D_i Z_i}{Z_{ERODED}}, & \text{otherwise}
\end{cases}$$  \hspace{1cm} (3.15)

The calculation of grain size changes due to deposition takes place at the conclusion of an event’s model run. The grain size of all sediment that has been deposited in a particular cell is either averaged with the grain size of the active layer (as in Equation 3.15, if the
amount of deposition does not require the addition of a new active layer), or is divided into as many new layers as are necessary given the user-defined layer thickness, each of which is assigned the averaged grain size of the deposited sediment. Note that while this routine may be used for the computation of grain size and stratigraphy at a particular cell, all model runs described here employed a single representative grain size across the model reach (see Section 3.5).

3.3 Model Validation

To compare the outputs of the model with field-based surveys of channel evolution, we derived several morphometric parameters along with comparing DEMs-of-Difference (DoDs)
and contributions of individual braiding mechanisms to total geomorphic change. Each of these three validation components are discussed below.

3.3.1 Morphometric Indices

We manually quantified braiding index \( I_B \) for the initial and final field surveys and model runs of each simulation described here. Braiding index was computed by averaging the number of channels across five evenly spaced transects along the length of the model domain (Howard et al., 1970; Egozi and Ashmore, 2008). Channels were defined by wetted areas as modeled using Delft3D at estimated baseflow for each modeling site. In addition, we measured the mode, or total sinuosity \( S_T \) of the modeled reach for the first and last model runs in each system. Total sinuosity (Richards, 1982) was defined by the ratio of the length of all anabranches \( L_A \) compared to the down-valley length of the model domain \( L_D \):

\[
S_T = \frac{L_A}{L_D}
\]

Finally, we computed the number of confluences, diffluences, and channel heads for initial and final field and model DEMs. The procedure for delineating confluences, diffluences, and channel heads is detailed by Wheaton et al. (2013). In brief, it requires manual location of areas where one anabranch splits into two anabranches (a diffuence), areas where two anabranches join to form one anabranch (a confluence), and locations where small side channels or chutes begin (a channel head). In theory, the number of confluences should be roughly equal to the number of diffluences plus the number of channel heads for a braided river; large differences in this metric are indicative of distributary systems (e.g., deltas; Jerolmack and Morhig, 2007) or dendritic networks that collect low order channels into a few main channels.

3.3.2 DEMs-of-Difference

We differenced DEMs of initial and final field surveys and model simulations using the Geomorphic Change Detection 6 software (http://gcd.joewheaton.org; Wheaton et al.,
Differencing DEMs from two survey periods produces a DEM-of-Difference (DoD), or a map of geomorphic changes that occurred during the inter-survey period. While the DEM differencing process is straightforward, accounting for error in the resultant DoD is necessary as each constituent DEM contains an inherent level of error (that may vary on a cell-by-cell basis), which can ultimately influence the estimated magnitude of geomorphic change in the DoD. Here we modeled DEM error using the most straightforward of the available approaches: we simply assumed that each of the constituent DEMs contained no error and computed the DoD. We then thresholded, or removed areas of change, from the DoD if they were less than 0.1 m in magnitude (herein termed the 'minimum level of detection' or minLoD). For decadal-scale modeling and field surveys on the River Feshie, we employed a threshold of ±0.2 m to better visualize and delineate braiding mechanisms. While this simple thresholding method is not necessarily the most robust available for modeling error in field-surveyed DEMs, those DEMs output from the model do not contain survey error as would be expected from field-based DEMs. As such, a simple minLoD of 0.10 m allowed us to use identical error modeling methods to directly compare areas of change in field and modeled DoDs while simultaneously removing a great deal of the survey noise/uncertainty present in field DEMs, along with removing extremely low-magnitude elevation changes in modeled DEMs.

From DoDs, we extracted elevation change distributions (a histogram of all volumetric changes), along with deriving the net sediment imbalance during each survey and model period (the percent departure of the sediment budget from equilibrium conditions).

3.3.3 Braiding Mechanisms

Using the GCD software, we mechanistically segregated thresholded DoDs by delineating the processes responsible for each area of geomorphic change (Ashmore, 1991; Wheaton et al., 2013). These processes can be separated into the four 'braiding mechanisms', or processes that act to reinforce or create a braided channel planform, described by Ashmore (1991): central bar development, lobe dissection, transverse bar dissection, and chute cutoff.
To these, Wheaton et al. (2013) added an additional six mechanisms that are not unique to braided rivers: bank erosion, channel incision (i.e. bed erosion), overbank sheets, confluence pool scour, bar trimming, and lateral bar development. Hereafter, these ten processes are referred to simply as 'braiding mechanisms'. We mapped the areas in DoDs where each of these mechanisms occurred; areas where we could not confidently assign a mechanism of change were classified as questionable or unresolved change. This classification allowed us to compute and compare the volumetric contribution of each braiding mechanism to total geomorphic change in both field and model-derived DoDs.

3.4 Modeling Sites

3.4.1 River Rees, South Island, New Zealand - Event and Annual Scales

The braided gravel bedded Rees (Figure 3.5) drains the uplifting metasedimentary Southern Alps and flows into Lake Wakitipu. The 2.5 km study reach is an actively braided channel which flows through a deglaciated valley, and the river is braiding in response to sediment delivery from the tectonically-active landscape (Williams et al., 2013). At the study reach, the Rees drains an area of 420 km². The hydrology of the system is dominated by response to glacial melt upstream, and undergoes floods in the spring, summer, and fall that may completely alter the morphology of the braidplain over the course of a single flood. A temporary gauging station at Invincible (4 km upstream of the study reach) operated from 2009-2011 and recorded a mean discharge of 19.8 m³/s with a maximum flow of 475 m³/s. Although the gauging station was located upstream of the study site, Brasington (personal communication) notes that flow attenuation between the two sites is negligible, and so herein the discharge at the Invincible gauging station is used to drive hydraulic components of the model. The survey data on the Rees include sub-meter digital elevation models (DEMs) constructed via terrestrial laser scanning (TLS; ground-based lidar) and real-time kinematic global positioning system (RTK-GPS) surveys. In total, ten floods ranging from 51 m³/s to 403 m³/s were captured as part of the ReesScan project (Brasington et al., 2012) from 2009-2011, with post-flood DEMs surveyed in the period between each high flow. These
pre- and post-flood DEMs, along with a continuous hydrologic record and a high degree of dynamism across the braidplain at the event scale, make the Rees an ideal candidate to examine the performance of the model at the event and annual scales. DEMs generated from field surveys on the Rees had a cell resolution of 2 m.

3.4.2 River Feshie, Scotland - Annual and Decadal Scales

The weakly-braided gravel bedded Feshie (Figure 3.5) is a tributary of the Spey River and drains 231 km$^2$ of mountainous, postglacial terrain. Underlain by metamorphic and igneous rocks, the basin ranges from around 230 m - 1260 m in elevation. The mean flow near the river’s outlet was reported by Ferguson and Werritty (1983) as $8 \text{ m}^3/\text{s}$ with $Q_5 = 80 \text{ m}^3/\text{s}$. Topographic data for the 1 km study reach of the Feshie consist of nine years of resurveys (2000, 2002-2008, 2013) comprising more than a decade of channel change using RTK-GPS (2000-2006) along with TLS and RTK-GPS fusion scans performed for three years (2007-8, 2013). Additionally, the Feshie dataset contains continuous hydrograph data ($\sim 55$ years) and aerial photo records ($\sim 60$ years), along with UK Ordnance Survey channel planform maps dating to 1869. The Feshie has been the site of a great deal of previous research ranging from bar morphodynamics (Ferguson and Werritty, 1983; Wheaton et al., 2013) to development of riverine survey and DEM-differencing/change detection methodologies (Brasington et al., 2007; Vericat et al., 2007; Hodge et al., 2009; Wheaton et al., 2010). The combination of annual resurveys capturing over a decade of channel change in combination with mapping and aerial photographs dating back over a century make the Feshie an ideal candidate to examine the performance of the model at the annual and decadal scales. In addition, the Feshie site provides a mechanistic contrast to the Rees in that overall flood-to-flood dynamism is reduced via vegetation cohesion and fine sediment (Ferguson and Werritty, 1983), and the dominant mechanisms of change vary from those seen on the Rees, particularly with regard to chute cutoff and bank erosion (Wheaton et al., 2013). DEMs generated from field surveys on the Feshie had a cell resolution of 1 m.

3.5 Results
3.5.1 Hydraulic Modeling Validation

3.5.1.1 River Rees

The use of Delft3D to model two-dimensional hydraulics in braided, gravel-bed rivers is discussed extensively by Williams et al. (2013), who specifically applied the hydraulic model to the River Rees. In general, the model is capable of reproducing field-observed velocities and inundation extents across the reach, although the use of 2D hydraulics requires increased computational time as compared to more simplified cross-sectional (e.g. 1D) approaches. The outputs of Delft3D on the Rees (depth, velocity, inundation extent) were compared with field-surveyed values across a range of discharges and used to calibrate the model parameters (Colebrook-White roughness and horizontal eddy viscosity) until good agreement was reached. The reader is referred to Williams et al. (2013) for more detailed analysis of the validity of Delft3D on the Rees (and in braided gravel-bed rivers in general). In short, here we used the same upstream and downstream boundaries as Williams et al. (2013) and kept both $k_s$ (0.10) and $\nu$ (0.10) constant in line with their results (Table 3.1).
Figure 3.5. Morphodynamic Modeling Sites. Overview maps of River Rees (left) and River Feshie (right). Hillshaded DEMs (0.5 m resolution for Rees and 1 m resolution for Feshie) are shown atop aerial photograph base. Information on data availability for both sites shown in lower panels.
Table 3.1. Delft3D Hydraulic Modeling Parameters

<table>
<thead>
<tr>
<th>Site</th>
<th>Sim. Time</th>
<th>Time Step</th>
<th>Cell Size</th>
<th>Nodes</th>
<th>$D_{50}$</th>
<th>W-C Roughness ($k_s$)</th>
<th>Eddy Viscosity ($\nu$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>River Rees</td>
<td>1 hour</td>
<td>0.025 min</td>
<td>2 m</td>
<td>352,231</td>
<td>0.035 m</td>
<td>0.10</td>
<td>0.1 m$^2$s$^{-1}$</td>
</tr>
<tr>
<td>River Feshie</td>
<td>1 hour</td>
<td>0.025 min</td>
<td>1 m</td>
<td>201,116</td>
<td>0.1 m</td>
<td>0.29</td>
<td>0.1 m$^2$s$^{-1}$</td>
</tr>
</tbody>
</table>
3.5.1.2 River Feshie Hydraulic Modeling and Validation

In contrast to the Rees, no comprehensive validation of Delft3D exists on the Feshie; as such, we leveraged existing surveys of wetted areas from 2003-2007 in concert with surveyed water depth in those years to examine the performance of Delft3D. Because field surveys were conducted at low flows to facilitate rapid measurement of braidplain topography, here we are only able to validate the results of Delft3D at these low flows. However, Delft3D has been employed and validated on gravel-bed braided rivers at flood stage (Javernick, 2013), demonstrating that the model can accurately reproduce flood-stage hydraulic features and can be used to drive morphodynamic evolution at the event-scale. For modeling on the Feshie, we estimated discharge by downscaling the average observed flow for the relevant survey period at the nearest gauging station (SEPA No. 8013, Feshie at Feshiebridge) located approximately 11 km downstream, using a coefficient of 0.71 (Wheaton et al., 2013).

We estimated the downstream water elevation using surveyed inundation extent in combination with the DEM for each year modeled. Downstream water surface estimated from the spatial data were cross-checked using a reach-scale conveyance calculation (Williams et al., 2013).

Results of our validation of Delft3D on the Feshie at low flow are shown in Figure 3.6. Here we report (a) the mean of depth differences between modeled and observed values ($D_{diff}$), along with (b) the congruence of the modeled and measured inundation extents ($F_c$; Bates and De Roo, 2000) as described by the ratio of intersection and union areal extents. These two metrics are described by equations 3.17-3.18, respectively.

$$D_{diff} = \frac{\sum_{i=1}^{n} x_{mod} - x_{obs}}{n}$$

(3.17)

$$F_c = \frac{IA_{obs} \cap IA_{mod}}{IA_{obs} \cup IA_{mod}} * 100$$

(3.18)

The validation metrics indicate that at low flow, Delft3D accurately predicted both depth and inundation extent across the Feshie study reach, as indicated by metrics comparing the two. Both $D_{diff}$ and $F_c$ are consistent with validation work performed by Williams et al.
(2013) on the Rees, and are indicative of good agreement between hydraulic model and field-observed flow characteristics. It is important here to note several factors that may cause our model results to differ slightly from field-observed values. First, because no measure of discharge was available at the study reach, we downscaled discharge at the nearest gauging station (by an empirically-derived coefficient of 0.71; Wheaton et al., 2013), and as such any errors in this downscaling coefficient would be propagated into our estimates of discharge. Second, the value of discharge taken at Feshiebridge was the average flow during that year’s survey period (on average two weeks’ time), and variability in discharge could lead to inaccuracy in our discharge estimates at the study reach along with small-scale inaccuracy of the field-observed inundation extent. Nevertheless, the agreement between observed and modeled depth and inundation extent across five years of surveys indicates that Delft3D is capable of predicting low-flow hydraulics at the Feshie study site, and in combination with the high-flow validation work of Williams et al. (2013), we argue that the model is well suited for driving the hydraulic component of MoRPHED at both high and low discharges on braided rivers.

3.5.2 Event-Scale Morphodynamic Modeling: River Rees

We modeled a single flood event on the River Rees that occurred between 8-16 December, 2009 as the result of heavy rainfall in the upstream watershed (Figure 3.7). Peak flows reached a maximum instantaneous discharge of 258.8 m$^3$/s at the upstream Invincible gauging station during the afternoon of 9 December. Two smaller peaks in the hydrograph of 75.0 m$^3$/s (afternoon of 8 December and morning of 12 December) also occurred during this flood. Our modeling employed a single representative grain size ($D_{50}$) of 0.02 m. Geomorphic change captured by pre- and post-flood terrestrial laser scanning revealed that the most volumetrically significant mechanisms of change were transverse bar conversion (22%), bank erosion (21%), and lobe dissection (19%). Qualitatively, event-scale dynamics across the study reach are marked by widespread geomorphic change, particularly in the center of the braidplain where development of a single main channel occurred via channel
incision, bank erosion, and avulsions of smaller anabranches leading to deposition and subsequent dissection of mid-channel bars. Geomorphic change on the edges of the braidplain was somewhat muted, consisting largely of infilling of anabranches and accretion of central bars. Elevation changes ranged from -1.70 m to +1.32 m (Figure 3.7). Braiding index ($I_B$) decreased from 3.6 to 2.0 following the flood, and total sinuosity ($S_T$) decreased from 4.5 to 2.9 (Figure 3.7).

Results of morphodynamic modeling on the Rees for this event are shown in Figure 3.7. Total model runtime for the single event was approximately 90 minutes. Elevation changes in the study reach ranged from -1.76 m to + 0.89 m. Overall, geomorphic change was concentrated near the center of the braidplain, similar to geomorphic change measured from field data. Large swaths of bank erosion along a central anabranch developed, although not to the extent seen in field data. In general, geomorphic change in the modeled DoD appears muted in comparison to the field-based DoD. This is reflected in the elevation change distribution (ECD) shown for field and model data on the Rees (Figure 3.7), particularly with regard to the erosional component of volumetric change (47,598 m$^3$ in the field compared to 20,663 m$^3$ in the model; Table 3.2). Similarly, depositional volumes were greater in the field (35,551 m$^3$) compared to those in the model (16,188 m$^3$; Table 3.2). Average magnitudes of erosion and deposition agree well between field and model results. On average, erosion depth across the study reach was 0.13 m as observed through field measurement, and 0.07 m when modeled. Deposition averaged 0.10 m in the field and 0.06 m in the model (Table 3.2). The most volumetrically-significant braiding mechanisms in this run were central bar development (25% of total volumetric change), bar edge trimming, (16%), and bank erosion (15%).

3.5.2.1 Hydrograph Discretization

The choice of model timestep is one of the more important considerations in morphodynamic modeling, whereby the user must strike the optimal balance between a model timestep fine enough to preserve computational stability, yet which is coarse enough to
allow computation over meaningful timescales (Brasington et al., 2007). To investigate the implications of altering the model timestep in MoRPHED, we discretized the hydrograph used in event-scale modeling on the Rees (Section 3.6.2) so as to model three discrete points over the course of the modeled flood (Figure 3.8). These discharges were 75 m$^3$/s, 254 m$^3$/s, and 75 m$^3$/s, respectively. Planform DoD, ECD, and volumetric contribution of individual braiding mechanisms from this discretized model run are shown in Figure 3.8. In general, morphologic changes were more widespread across the braidplain in the case of the discretized hydrograph modeling run (Figure 3.8), with overall volume of change increasing, yet still smaller than field-observed change volumes (381,336 m$^3$ in the model compared to 421,468 m$^3$ in field data; Figure 3.8, Table 3.2). The ECD from this model run more closely approximated the field-derived ECD, with low-magnitude elevation changes (e.g. < 1 m) dominating the change distribution (Figure 3.8).

3.5.3 Annual-Scale Morphodynamic Modeling: River Feshie

We modeled morphodynamics during the one-year period from July 2003 to July 2004 on the Feshie. The estimated hydrograph for the study reach, based on the empirical downscaling coefficient applied to the Feshiebridge gauge (Section 3.3.1) during this period is shown in Figure 3.9. This survey epoch contained 16 flood peaks above the 'low bankfull' discharge estimate (20 m$^3$/s; Ferguson and Werritty, 1983) at the study reach; these are denoted in Figure 3.9 and were used as model inputs. While the use of bankfull discharge as an estimate of flow competence is simplistic, we use this threshold for two reasons: (a) Wheaton et al. (2013) noted that the number of flow peaks in excess of bankfull discharge in any year was related to the amount of geomorphic change over a period of five years, and (b) Ashworth and Ferguson, (1989) documented that flows of $\sim$22 m$^3$/s were indeed competent for bed material in the study reach, albeit without full mobility of all bed particle sizes. As such, the use of a low bankfull estimate for competence provides an easily measured threshold for competent flows that we know is related to geomorphic change in the study reach. Our modeling on the Feshie employed a single representative grain size (D$_{50}$) of 0.05
Wheaton et al. (2013) note that the most volumetrically significant braiding mechanisms during the 2003-2004 period were chute cutoff (29%), bank erosion (16%), and channel incision (15%). Overall geomorphic change was primarily confined to a main channel bisecting the braidplain longitudinally, with one anabranch on the left side of the braidplain undergoing bank erosion and central bar development. Braiding index ($I_B$) during the 2003-2004 epoch increased from 1.8 to 2.8, and total sinuosity ($S_T$) also increased from 2.5 to 3.5. Results of morphodynamic modeling are shown in Figure 3.9. Total model runtime for the series of 12 events modeled during the 2003-2004 period was approximately 6 hours.

The results of DEM differencing are shown in Table 3.3. Elevation changes in the modeled reach ranged from -1.25 m to +1.49 m. Overall geomorphic change was marked by the accumulation of transverse bars and the development of central bars throughout the model reach, along with incision of a central anabranch. Sculpting or trimming of central bars (Figure 3.9; e.g. Wheaton et al., 2013) was also prevalent. The most volumetrically-significant braiding mechanisms during this model run were channel incision (26%), transverse bar conversion (20%), and central bar development (14%). Overall, geomorphic change predicted via modeling was greater than that observed from field data (Table 3.3). However, both field and model ECDs (Figures 3.9.B, 3.9.C) depict change distributions wherein the greatest volume of geomorphic change is the result of low-magnitude elevation changes. The average depth of elevation changes were generally well predicted by the model, although average erosion and deposition depths were over-estimated by 0.06 and 0.03 m, respectively (Table 3.3).

3.5.3.1 Case Study: Contrasting Path-Length Distributions

Transport and deposition of eroded bed or bank sediment in MoRPHEd is a function of the path length distribution used in the model. While field or laboratory data describing particle transport distances can be used to produce a path length distribution, parameterizing such a distribution for sites where tracer data are not available is not straightforward.
Here we employed average confluence-diffuence spacing to estimate the peak of the distribution (Figure 3.3; Pyrce and Ashmore, 2003a,b; Kasprak et al., 2015). To further understand the effect of contrasting path length distribution shapes and distances, we used MoRPHED to model the annual hydrograph on the River Feshie in a manner identical to Section 3.3.1, except we varied the characteristics of the specified path length distribution (Figure 3.10). We employed a compressed Gaussian distribution (Figure 3.10.A), a flattened Gaussian distribution (Figure 3.10.B), and an exponential decay-type distribution (Figure 3.10.C; Pyrce and Ashmore, 2003b); each of these distributions had a total length identical to our original distribution on the Feshie (Figure 3.3.A). We also modeled two shortened distributions (length = 50 m): a shortened Gaussian distribution (Figure 3.10.D) and a shortened exponential distribution (Figure 3.10.E).
Figure 3.6. River Feshie Hydraulic Modeling. Surveyed water surface extent in each of five years (2003-2007) was compared to modeled inundation extent in the same years using discharge levels approximated using data from gauge at Feshiebridge (see Section 3.5.1). Areas observed (but not predicted) to be inundated shown in blue, areas predicted (but not observed) to be inundated shown in green. Areas which were correctly predicted as being inundated shown in red. Base is hillshaded 1 m DEM.
Figure 3.7. River Rees Event Morphodynamic Modeling Results. Continuous hydrograph and modeled discharge shown in (A). Elevation change distributions derived from field and model data shown in (B) and (C) respectively. DoD derived from field data shown in (D), with model-derived DoD and delineated braiding mechanisms shown in (E) and (F). Volumetric contributions of each braiding mechanism in field and modeled data shown in (G).
Table 3.2. River Rees Event Modeling: Geomorphic Change Detection Results; results of discretized modeling shown in *italics*

<table>
<thead>
<tr>
<th></th>
<th>Field Raw</th>
<th>Field Thresholded (0.1 m)</th>
<th>MoRPHED Raw</th>
<th>MoRPHED Thresholded (0.1 m)</th>
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<td><strong>Areal</strong></td>
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<tr>
<td>Total Area of Erosion (m²)</td>
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<td></td>
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<tr>
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<tr>
<td>Average Depth of Erosion (m)</td>
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<tr>
<td>Average Depth of Deposition (m)</td>
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<td>0.06</td>
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<td><em>0.07</em></td>
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<td>Average Total Thickness of Difference (m)</td>
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<td><em>0.00</em></td>
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Figure 3.8. River Rees Discretized Hydrograph Case Study. DoD from MoRPHED modeling shown in (A), with ECD in (B) and event hydrograph with model points shown in (C). Refer to DoD and ECD in Figure 3.7 for comparison.
Figure 3.9. River Feshie Annual Morphodynamic Modeling Results. Refer to Figure 3.7 caption for details.
Table 3.3. River Feshie Annual Modeling: Geomorphic Change Detection Results

<table>
<thead>
<tr>
<th></th>
<th>Field Raw</th>
<th>Field Thresholded (0.1 m)</th>
<th>MoRPHED Raw</th>
<th>MoRPHED Thresholded (0.1 m)</th>
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<td><strong>Areal</strong></td>
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<td></td>
<td></td>
<td></td>
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<tr>
<td>Total Area of Erosion (m²)</td>
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<td>7,236</td>
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<tr>
<td>Total Volume of Erosion (m³)</td>
<td>4,433</td>
<td>2,806</td>
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<td>5,239</td>
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<td>Total Volume of Deposition (m³)</td>
<td>2,704</td>
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<td>5,986</td>
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<tr>
<td>Average Depth of Erosion (m)</td>
<td>0.07</td>
<td>0.28</td>
<td>0.12</td>
<td>0.30</td>
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<tr>
<td>Average Depth of Deposition (m)</td>
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<td>0.21</td>
<td>0.10</td>
<td>0.24</td>
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<td>Average Total Thickness of Difference (m)</td>
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<td>0.04</td>
<td>0.11</td>
<td>0.09</td>
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<td>Average Net Thickness of Difference (m)</td>
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<td>-0.01</td>
<td>0.00</td>
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Figure 3.10. River Feshie Variable Path Length Case Study. Path length distributions used are: compressed Gaussian (A), stretched Gaussian (B), and exponential (C), along with shortened Gaussian (D) and shortened exponential (E). DoDs for each path length distribution shown in middle row, with elevation change distributions shown in bottom row.
Overall, the geomorphic changes predicted by the model were strikingly similar at the reach scale, and areas of scour and deposition generally aligned between all five of the simulations (Figures 3.10.A - 3.10.E). Analysis of the elevation change distribution produced using each path length distribution revealed that while the compressed and stretched Gaussian distributions were marked by more laterally-extensive deposition at higher magnitudes (e.g. $\sim 1$ m), the exponential and two shortened distributions (Figures 3.10.C - 3.10.E) generally contained depositional signatures marked by numerous low-magnitude (e.g. $\Delta z < 0.5$ m) changes. The same is true for the erosional component of elevation change, with compressed and stretched Gaussian distributions marked by a wide range of erosional elevation changes up to and exceeding 1 m, whereas the exponential and shortened distributions generally displayed erosional changes less than -1 m in elevation.

3.5.3.2 Decadal-Scale Morphodynamic Modeling: River Feshie

We modeled morphodynamics during the ten-year period between July 2003 and June 2013 along the River Feshie. The estimated hydrograph at Glen Feshie during this period is shown in Figure 3.11.A, and we modeled all peaks above the low bankfull discharge as described in Section 3.5.2 and in Wheaton et al. (2013), for a total of 185 flood events ranging from $20 \text{ m}^3/\text{s}$ to $95 \text{ m}^3/\text{s}$.

Differencing DEMs from survey data at the beginning and end of the analysis period reveals elevation changes ranging from 2.4 m to $+2.1$ m (Figure 3.11.D). DEM differencing indicates that the study reach underwent slight net aggradation ($+3.6\%$ imbalance). The most volumetrically significant braiding mechanisms during this time period were the development of central bars (25% of volumetric changes), transverse bar conversion (17%), and bank erosion (16%). As the relative contribution of individual braiding mechanisms may be misleading at decadal scales due to signature overprinting and hence difficulty in interpretation of braiding mechanism (Collins et al., 2012), we note that over a 5-year period (2003-2007; Wheaton et al., 2013) on the Feshie, the most volumetrically significant braiding mechanisms were chute cutoff (24%), bank erosion (20%), and transverse bar con-
version (19%). Braiding index ($I_B$) during the 2003-2013 epoch increased from 1.8 to 2.4, and total sinuosity ($S_T$) also increased from 2.5 to 2.9 (Figure 3.11).

Results of morphodynamic modeling from 2003-2014 are shown in Figure 3.11. Total model runtime for the 185-flood series was approximately 72 hours. Geomorphic change ranged from +2.20 m to -5.96 m (Figure 3.11.E). A mask was employed to exclude areas of the reach < 25 m from the downstream boundary and < 50 m from the upstream boundary, as boundary artifacts in these areas resulted in high magnitudes of geomorphic change (e.g. > 5 m m) that were attributable to hydraulic artifacts (i.e. deep scour at discharge points used in Delft modeling). While geomorphic change in the field was marked by generally thin-mantled erosion and deposition across the braidplain, the model produced more widespread, high-magnitude erosional change (Table 3.4). While the ECD produced by the model generally characterized the form of the field-derived ECD (Figures 3.11.B, 3.11.C), the model predicted a smaller area, but greater volume, of scour (Table 3.4). The model generally reproduced the form and magnitudes of deposition seen in the field, but the lowest-magnitude deposition (e.g. < 0.5 m) was more volumetrically significant in the field than in the model. In particular, the model did not produce avulsions seen in the field, but rather predicted continued downcutting of the central anabranch seen in the 2003 DEM. The most volumetrically-significant braiding mechanisms during the 2003-2013 model run were channel incision (32%), central bar development (18%) and transverse bar conversion (18%). In addition, the role of bar edge trimming (8%), a process treated identically to bank erosion in MoRPHED, was magnified compared to field-derived mechanistic segregation (2%).
Figure 3.11. River Feshie Decadal Morphodynamic Modeling Results. Refer to Figure 3.7 caption for details.
Table 3.4. River Feshie Decadal Modeling: Geomorphic Change Detection Results

<table>
<thead>
<tr>
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<th>Field Raw</th>
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<th>MoRPHED Thresholded (0.2 m)</th>
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<tr>
<td>Total Area of Erosion (m²)</td>
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<td></td>
</tr>
<tr>
<td>Total Volume of Erosion (m³)</td>
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<td><strong>Vertical Averages</strong></td>
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<tr>
<td>Average Depth of Erosion (m)</td>
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<td>0.80</td>
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<td>Average Depth of Deposition (m)</td>
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<td>0.41</td>
<td>0.26</td>
<td>0.51</td>
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<tr>
<td>Average Total Thickness of Difference (m)</td>
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<td>0.25</td>
<td>0.32</td>
<td>0.28</td>
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<tr>
<td>Average Net Thickness of Difference (m)</td>
<td>0.04</td>
<td>0.02</td>
<td>-0.01</td>
<td>-0.02</td>
</tr>
</tbody>
</table>
3.6 Discussion

We developed a morphodynamic model that computes sediment transport according to user-specified path length distributions, and subsequently employed this model to predict channel evolution at two braided river reaches across timescales ranging from a single event to a decade. We observed that the model reproduced many of the geomorphic changes observed in the field, although the magnitude and mechanisms of those changes were often poorly predicted. At the same time, the modular design of the modeling framework may hold promise for exploration of braided channel evolution, and also raises questions regarding the way processes are represented algorithmically and the model’s sensitivity to those process representations.

3.6.1 Emergent versus Parameterized Processes

Braided rivers undergo geomorphic change as a result of numerous morphodynamic processes, or braiding mechanisms (Ashmore, 1991; Wheaton et al., 2013; Kasprak et al., 2015). Given its highly simplified nature, the degree to which these braiding mechanisms must be explicitly represented as algorithms in morphodynamic modeling deserves exploration. Of the ten braiding mechanisms discussed in Section 3.4.3, the model produced eight simply as a result of the bed erosion, transport, and deposition functions included in the model. Only bank erosion and bar edge trimming required the inclusion of a lateral channel migration component. As the surfaces undergoing bank erosion and bar edge trimming are not necessarily submerged during a flood, these processes cannot be captured simply by an excess shear stress scour approach (Equation 3.8; Section 3.3.2).

The geomorphic changes that our model most commonly produces are those that result from focused scour and longitudinally-continuous deposition, given the nature of the scour and deposition functions used in the model (Sections 3.3.2, 3.3.3). In particular, channel incision, bar edge trimming, bank erosion, and lateral/central bar development are common processes produced by the model (Figures 3.7, 3.9, 3.11). Additionally, given the single-peaked Gaussian distributions used herein, those braiding mechanisms which in-
volve deposition immediately downstream of an erosional source are difficult to reproduce. For example, Wheaton et al. (2013) demonstrated the importance of chute cutoff as a braiding mechanism on the Feshie, noting that chute development across point bars not only manifested as erosion, but that the scoured material was often deposited immediately downstream of the chute. Similarly, scoured bank material (e.g. mass failures) may often be deposited at the bank toe rather than transported downstream. In both cases, the model makes no differentiation in transporting the eroded sediment, mobilizing the material according to the user-specified path-length distribution; as such, proximal couplets of erosion and deposition are difficult to reproduce in the model. Finally, we note that chute cutoff in the model always occurred in locations of pre-existing chutes across point bars. Because chute cutoff often occurs at the falling stage of floods, when braidplain-inundating flows are first being confined into anabranches, our model may not properly reproduce chute cutoff as a result of only computing peak flood hydraulics, and averaging shear stress across a range of model cells. Although headward erosion of these pre-existing chutes typically occurred in the model, thus increasing their extent, we did not observe any instances where chute cutoff was initiated in the model without an existing chute or channel head being present on a bar surface. As such, chute cutoff may represent a braiding mechanism that must be explicitly included in our model’s code in order to be properly represented in the future.

3.6.2 Sensitivity to Process Representation

3.6.2.1 Hydrograph Discretization

In Section 3.6.2, we modeled a single event on the Rees as three discrete discharges on the hydrograph (Figure 3.8). Because MoRPHED under-predicted the volume of change, particularly due to the absence of low-magnitude scour, during the single-event simulation (Figure 3.7), we sought to understand whether discretizing the hydrograph would allow for an improved prediction of overall volumetric change, and low-magnitude erosional change in particular. Overall, discretizing the hydrograph into three modeling timesteps only marginally increased predictions of volumetric change across the study reach: total
volumetric change in the discretized run (Figure 3.8) was 39,771 m$^3$, compared to 36,850 m$^3$ in the single-event model run (Figure 3.7). It might appear counterintuitive that discretizing the hydrograph and running the model three times would not appreciably increase the amount of volumetric change. We believe the reason for this is that in many cases, deposition offsets erosion, particularly in inundated areas, leading to little net increase in volumetric change despite running the model multiple times. The exception to this is in areas that experience bank erosion or avulsions, where high-magnitude scour may occur in areas which are not inundated, thereby producing locations where deposition does not counteract scour. Both modeling approaches underestimated the amount of volumetric change in the field (83,149 m$^3$).

Discretizing the hydrograph also increased the amount of low-magnitude scour predicted by the model (Figure 3.8.B), more accurately reflecting the field-derived elevation change distribution (Figure 3.8.C). It is likely that this is the result of the 0.1 m elevation thresholding used in our change detection (Section 3.4.2), whereby additive changes due to erosion largely did not exceed 0.1 m depth after a single flood event, but did exceed this threshold when three discrete hydrograph points were measured. Whereas erosional processes that lead to high-magnitude scour, such as bank erosion, dominated the elevation change distribution in the single-event model run, processes such as channel incision and lobe dissection were more prevalent in the discretized hydrograph model run. Additionally, several areas of high-magnitude bank and bar trimming were largely offset by deposition of imported or scoured material during the discretized hydrograph run, thereby decreasing the overall magnitude of scour in those areas (Figure 3.8.B).

3.6.2.2 Path Length Distribution

We modeled annual-scale morphodynamics on the Feshie using five different path length distributions (Section 3.6.3). The similarity between the DEMs produced by the model (and hence, the DoDs shown in Figure 3.10) using these distributions is striking, yet subtle differences and may reflect the distinct nature of the spatial arrangement of erosion and
deposition. Overall, the similarity between the modeled distributions may also be the result of the smoothing algorithms used in the model to ensure computationally-stable output surfaces, such as the along-flow averaging of shear stress and neighborhood windows used for deposition (Section 3.3.2), both of which may act to reduce the variability introduced by the choice of a particular path length distribution.

In fluvial settings, erosional processes typically operate over small spatial scales (e.g. bank erosion, bar trimming, pool scour), and the magnitude of scour in these focused areas is typically higher than diffuse, broad-scale depositional processes such as overbank sheets or accretion of mid-channel or lateral bar material (Wheaton et al., 2013). As such, we suggest that the overall similarities, as well as the differences between our model’s outputs using these contrasting path length distributions reflects the ability of deposition to counterbalance elevation changes due to scour of material, and the fact that the diffuse nature of the path length distributions used here make this counterbalancing difficult. For example, the compressed and stretched Gaussian distributions (Figures 3.10.A, 3.10.B) are both marked by broad areas of erosion and deposition typically falling between ± 1 m in elevation change. However, high-magnitude areas of erosion are more rare, yet still present, in the stretched Gaussian distribution, which may be due to the more longitudinally-extensive deposition of scoured material partially offsetting elevation changes due to erosion. The compressed and stretched Gaussian distributions stand in contrast to the exponential and shortened distributions (Figures 3.10.C - 3.10.E), where elevation changes are largely confined between ± 0.5 m, and overall are more fragmented across the model reach. This does not reflect a reduced magnitude of erosion, as Equation 3.8 was used in all cases to predict scour depth; rather, the fragmented nature of elevation changes, along with the narrower range of those changes, is likely due to the propensity for deposition to offset erosional changes given the more focused nature of the path length distributions in Figures 3.10.C - 3.10.E. Nevertheless, in all distributions used, the volume of material deposited in a given cell following erosion upstream is always a fraction of that which was eroded. Using the distributions in Figure 3.10 and the deposition smoothing window detailed in Section 3.3.3,
the volume of sediment deposited falls between 0.4% and 8% of the volume which is eroded, and as such, erosion may outpace deposition in many cells.

3.6.3 Imperfect Models as Exploratory Tools

Models in the earth sciences are necessarily imperfect (Oreskes et al., 1994), and the highly simplified nature of MoRPHED, combined with the highly dynamic and nonlinear nature of braided river morphodynamics (Ashmore, 1991; Bristow and Best, 1993) implies that our model will necessarily fail to achieve perfect simulation of field-observed geomorphic dynamics. However, even imperfect models can provide meaningful insight into the processes behind morphologic evolution of fluvial systems (Paola et al., 2009; Paola and Voller, 2009). MoRPHED is designed to facilitate experimentation, particularly with regard to process inclusion or the particular aspects of process representation of bed and bank erosion, transport, deposition, and import dynamics (Figure 3.1). For example, in Section 3.6.3, we explored the implications of altering the path-length distributions for bed and bank sediment transport/deposition, along with seeking to understand the advantages and drawbacks of discretizing hydrographs during model runs. The notion of morphodynamic modeling that employs sediment transport routines based on particle path length distributions is in its infancy, and we have built the model as an exploratory tool that can be used to investigate the utility of this approach toward predictive modeling of braided river evolution. Several components of the model may be employed to investigate longstanding questions in our understanding of braided river dynamics, starting with the path-length approach itself. While it has long been hypothesized, and field and laboratory data have often confirmed, that mobilized particles in braided rivers are preferentially deposited in association with regularly-occurring channel bars (e.g. Pyrce and Ashmore, 2003a,b; Kasprak et al., 2015), the form of the path length distribution, and its relationship to geomorphic unit spacing, is deserving of further study across braided systems. As such, the choice of path length distribution and subsequent comparison of model results with field observations may provide insight into the applicability of path length distributions on a system-by-system
MoRPHED may also be used to investigate the utility of event-scale monitoring. Existing morphodynamic models typically employ a sediment continuity approach (e.g. Exner; Equation 3.4), operating at very fine temporal scales, typically seconds to minutes. This approach produces results consistent with field observation (Bates et al., 2005), but comes at the expense of computational overhead, thereby restricting the timescales that can be modeled (Brasington et al., 2007). Because the timestep of the model is, by default, a single event, computational resources are conserved, allowing for extended simulations at annual and decadal timescales. However, it is unclear whether processes that occur over the course of a competent flow (e.g. avulsions, bank failures) can be adequately captured using an event-scale modeling approach. Similarly, the degree to which a hydrograph may need to be discretized and its constituent parts modeled (Section 3.6.2) in order to capture stage dependent processes such as the development of chute cutoffs or bar edge trimming (Wheaton et al., 2013) is deserving of further investigation.

Our exploratory research into decadal-scale modeling on the River Feshie (Section 3.6.4) indicates that both the accurate prediction of event-scale scour depth, and subsequent deposition location, present significant methodological hurdles in the development of valid morphodynamic models that operate at the timescale of competent flows. In our model, as in the field, erosion occurs in discrete, focused areas of high magnitude (e.g. pool scour, bar trimming, bank retreat; Ashmore, 1991; Bristow and Best, 2003; Wheaton et al., 2013). Deposition occurs thinly over more broad spatial areas (e.g. bedload sheets, overbank sheets, bar formation), a result of the path length distributions employed here and the smoothing required to avoid the generation of rough topography that would lead to hydraulic instability. One unfortunate result of the differences in the nature of erosion and deposition is that deposition may never 'catch up' to erosion if a simple path-length distribution is employed, thus resulting in over-scouring of channels (Figure 3.11). In general, the morphodynamic signature of deposition mirrors that seen in the field (see ECDs in Figures 3.7, 3.9, 3.11), with a large contribution of total change coming from areas of shallow
deposition. However, in future event-scale morphodynamic models, it may be necessary to augment path length distributions so as to preferentially deposit material in certain geomorphic units (e.g. confluence pools, deep channels adjacent to banks) in order to develop lateral flow, bank material removal, and channel migration/avulsion (Ashmore, 1991).

Another process that has proved troublesome in the development of MoRPHED is the lateral retreat of banks. Highly erodible banks are a hallmark of braided rivers and lead to the development of central bars and multiple anabranches (Ashmore, 1991). However, the Cartesian grid employed in the model, along with the event-scale timestep of the model, makes the generation of smooth bank features difficult. While approaches are available that compute bank stability based on a factor-of-safety approach that balances downslope gravitational forces with the ability of bank material to provide cohesive resistance to failure (Darby and Thorne, 1996; Simon et al., 2000; Rinaldi and Darby, 2007), parameterization of these models, especially at the reach scale, is quite difficult. The simplified approach employed in the model averages the slope of bank cells and the near-bank shear stress of the flow to predict bank retreat distances (Section 3.3.4). The threshold slope and shear stress, and their effect on lateral retreat distance, have been empirically adjusted to emulate field-observed bank dynamics. We have observed that the simple treatment of lateral erosion in the model produces bank erosion and bar edge trimming. However, whether this simplified approach will provide computational stability over longer-term (e.g. centennial) simulations is unknown. Additionally, further investigation is needed to determine whether the inclusion of bank toe deposition, as opposed to immediate downstream transport according to a user-specified path length distribution, is necessary in the model, along with whether bank material should be transported and deposited according to the same path-length distribution as bed material (Section 3.3.3).

3.7 Conclusions

The developed morphodynamic model has been developed to simulate braided river
evolution at a variety of timescales via a path-length based approach, and here we applied
the model to two braided river reaches at the event, annual, and decadal scale. The premise
of MoRPHED is that particle travel distances can be approximated using path length dis-
tributions (Pyrce and Ashmore, 2003a,b; Kasprak et al., 2015), which results in decreased
computational overhead when modeling sediment transport, thus enabling longer simulations
at improved spatial resolution. We observed overall correspondence between model
outputs and field observations when comparing planform changes, geomorphic change de-
scribed by elevation change distributions, and morphometric indices such as sinuosity, braid-
ing index, and channel node counts (Section 3.4). At the same time, divergence between
field and model datasets was evident, suggesting that a purely path-length-based approach
may oversimplify the highly dynamic nature of braided systems (Bristow and Best, 1993).
While we did observe reproduction of all field-observed braiding mechanisms (Wheaton et
al., 2013), the relative contribution of these mechanisms often varied from values seen in
the field. In contrast to all other braiding mechanisms, neither bank erosion nor bar edge
trimming emerged simply as a result of bed scour and deposition, and instead needed to
be explicitly parameterized in the model. In addition, chute cutoff only occurred at areas
where pre-existing chutes or channel heads were observed, and did not appear to emerge
across previously flat bar tops. While this model represents a first step in event-scale, path-
length-based morphodynamic modeling, it remains to be seen whether the approach will
prove feasible for longer-term modeling runs and/or whether process representation that
accounts for inter-flood geomorphic change, such as avulsions or bank mass failure (Leddy
et al., 1993; Ashworth et al., 2004) will require explicit parameterization. Perhaps more
importantly, we argue that MoRPHED should not be judged as a perfect model, as such
a task is impossible (Oreskes et al., 1994). On the contrary, we have designed the model
to be a modular framework for exploring the effect of various process representations, and
their inclusion or exclusion from the model, as a learning tool designed to reveal the rel-
ative importance of geomorphic transport processes in braided river dynamics at multiple
timescales.
3.8 Acknowledgments

This research was supported by a grant from the National Science Foundation (No. 1147942). We thank Philip Bailey (North Arrow Research) for extensive assistance with model code and algorithm development. Field work on the Feshie was completed with the generous permission of Thomas MacDowell and the Glenfeshie Estate, with the assistance of Niall Lehane (Queen Mary University of London), Mark Smith (University of Leeds), Julian Leyland (Southampton University), and Damia Vericat (Forest Technology Institute of Catalunya). Our modeling efforts benefited from substantial discussion with Sara Bangen, Nate Hough-Snee, Wally MacFarlane, Eric Wall, and Peter Wilcock (Utah State University), along with Rebecca Hodge (Durham University), David Sear (Southampton University), and Richard Williams (University of Aberystwyth).
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CHAPTER 4

THE SENSITIVITY OF BRAIDED RIVER MORPHODYNAMICS
TO VARIATIONS IN SEDIMENT SUPPLY
AND SOURCE

Abstract

In braided channels, the transient nature of bars points to a readily-available supply of sediment necessary for creating and maintaining the braided planform. Numerous studies have documented planform changes in braided rivers resulting from climate or land use activities that alter sediment delivery regimes, whereby transitions to or from single-thread channels may result from a decrease or increase in upstream sediment supply, respectively. Braided rivers are often found in valleys floored by large volumes of sediment (e.g., in proglacial settings or recently deglaciated valleys), suggesting that ample sediment for bar building often exists within a given reach. This may imply that braided rivers are inherently insensitive to alterations in external sediment delivery regime for some time. Here we use scenario-based morphodynamic modeling to investigate the geomorphic response of a reach of the wandering River Feshie, Scotland, to variations in sediment supply from upstream. Four scenarios were modeled, describing sediment equilibrium, deficit, surplus, and quasi-equilibrium marked by periodic import of sediment. Geomorphic change was quantified by computing areal and volumetric changes across the reach, along with measuring the relative volumetric contribution of eleven braiding mechanisms. Of the four modeled scenarios, only the equilibrium and quasi-equilibrium simulations maintained the braided nature of the modeled reach after a ten-year period. The relative contribution of braiding mechanisms was similar between these two simulations, and the most volumetrically-significant mechanisms were channel incision, lobe dissection, and central bar development. The modeled scenarios produced changes that were of a greater magnitude and confined to a smaller area of the braidplain than changes observed in the field. Nevertheless our scenario-based

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1This chapter is in preparation for submission to Geology and is co-authored with Konrad Hafen and Joseph Wheaton (Utah State University) and James Brasington (Queen Mary, University of London).
modeling indicates that while relative sediment equilibrium is necessary for development and maintenance of the braided planform, the frequency with which sediment is imported from upstream may vary significantly, and during decadal periods when upstream supply is limited, local sediment sources are sufficient for bar building and planform maintenance.

4.1 Introduction

High sediment supply is a prerequisite for the formation and maintenance of braided rivers (Ashmore, 1991; Leopold and Wolman, 1957; Mueller and Pitlick, 2014), together with the competence to mobilize bed material and readily erodible banks that lead to deposition and reworking of bed sediment. Supplied sediment is the source of material for building a network of transient bars, the hallmark of the braided planform, around which channels split and rejoin (Brierley and Fryirs, 2005; Bristow and Best, 1993). Over more than six decades of research, the morphodynamic processes giving rise to the braided planform have been studied using field and laboratory research, along with numerical modeling.

Despite the fundamental importance of sediment supply in the development of braided channels, the issue of sediment source remains almost completely unexplored. On one hand, the supply of sediment for building bars and developing the braided planform may intuitively be sourced from upstream, as numerous studies focusing on braided to single-thread transitions following elimination of upstream supply (e.g., dams) have demonstrated (Marston et al., 1995; Gilvear, 2004; Gurnell et al., 2009). Yet to solely focus on upstream supply ignores the fact that braided rivers are commonly found in valleys floored by sediment available locally for reworking, such as those braided streams in deglaciated valleys or downstream from glaciers (Goff and Ashmore, 1994; Wheaton et al., 2013). These cases provide evidence that the dominant source of sediment for bar building and establishment of the braided planform may lie within the reach itself, and be accessed through channel migration and avulsion. This notion is reinforced by field studies that have shown rapid planform transitions in braided reaches when sediment was locally removed via gravel mining (Piégay et al., 1997; Kondolf et al., 2002) or channel control measures that restrict access
to sediment sources on the braidplain (Roux et al., 1989).

The question of upstream sediment supply versus within-reach sediment supply to bars in braided rivers has tremendous implications for our understanding of these systems. For example, shifting climatic regimes that may alter the relative frequency of sediment-supplying floods to braided reaches have the potential to force reconfiguration of the braided planform if upstream sediment supply is the predominant source of bar-building material (Winterbottom, 2000; Vandenbergh, 2002; Vandenbergh, 2003). Conversely, if most of the material necessary for bar building and braided planform maintenance has its source within the reach, braided channels may be buffered from altered flood and sediment supply regimes. The same questions arise when considering the effect of any anthropogenic sediment supply alterations to braided rivers, such as dams or shifts in land use that change the timing or magnitude of sediment delivery to channels (Beguería et al., 2006; Kasprak et al., 2013).

One potential explanation for the persistence of this knowledge gap is attributable to the extended timescales over which channels respond to altered sediment supply, often rendering direct field observation intractable (Schmidt and Wilcock, 2008; Gurnell et al., 2009). Numerical models that compute channel adjustments to water and sediment supply (e.g., morphodynamic models) provide a way forward for understanding and predicting the effect of altered sediment regime on channel morphology. At the same time, the high computational overhead, or conversely, the physical simplifications necessary for modeling at relevant spatiotemporal scales (e.g., annual to centennial over reaches ranging from 1 - 10 km) mean existing morphodynamic models fail to predict the morphologic impact of altered sediment supply on rivers (Brasington and Richards, 2007).

4.1.1 Objectives

To address this knowledge gap, here we use scenario-based morphodynamic modeling to explore the effect of altered sediment supply on channel morphology at annual to decadal timescales. We seek to quantify how the morphodynamic processes that give rise to - and maintain - braided channel planforms vary based on the amount of upstream sediment
supplied to a reach. We use simple scenario-based numerical modeling to predict the morphodynamic response of a wandering gravel-bed river to altered sediment supply regimes over decadal timescales.

4.2 Study Setting

A 1-kilometer reach of the gravel-bed River Feshie, Scotland is used as the modeling site for this research. The weakly-braided Feshie is a tributary of the Spey River and drains 231 km$^2$ of mountainous, once-glaciated terrain. Underlain by metamorphic and igneous rocks, the basin ranges from around 230 m - 1260 m in elevation. The mean flow near the river's outlet is about 8 m$^3$ sec$^{-1}$, with Q$_5$ (the five-year recurrence interval flood) equal to 80 m$^3$ sec$^{-1}$ (Ferguson and Werritty, 1983). Topographic data for the study reach of the Feshie, herein termed the Glenfeshie Reach, consist of nine years of annual resurveys (2000, 2002-2008, 2013) using real-time kinematic global positioning system (rtk-GPS) and terrestrial laser scanning (TLS, aka ground-based lidar). For all model simulations herein, we simulated channel evolution during the period 2003-2013, for which continuous 15-minute hydrograph data are available from the Scottish Environmental Protection Agency (SEPA; Station 8013, Feshie at Feshiebridge, 11 km downstream of the study reach).

Wheaton et al. (2013) discuss the development of an empirical downscaling coefficient (0.71), which we employed here to translate SEPA-reported discharges at Feshiebridge to those at Glenfeshie. While we did not perform direct measurements of competent discharge at the study reach, Ferguson and Werritty (1983) noted that competent flows are roughly 20 m$^3$/s in the Glenfeshie Reach. The Feshie has a rich history of studies seeking to understand bar evolution (Ferguson and Werritty, 1983; Wheaton et al., 2013), sediment transport (Ashworth and Ferguson, 1989; Hodge et al., 2011), and the development of modeling and monitoring techniques for braided rivers (Brasington et al., 2007; Vericat et al., 2007; Wheaton et al., 2010; Brasington et al., 2012; Rychkov et al., 2012).
4.3 The Model and Modeled Scenarios

We employed the newly-developed MoRPHED morphodynamic model (see Chapter 3) that uses simplified sediment transport routines based on characteristic travel distances, or particle path lengths to computed sediment transport and resultant channel morphology at the scale of individual flood events over reach (e.g., kilometer) scales. Full documentation of the operation of MoRPHED and its components can be found in Chapter 3 and at http://morphed.joewheaton.org. Here we briefly discuss the model’s operation and requisite input data, which are also detailed in Figure 4.1.
MoRPHED operates on a user-input digital elevation model (DEM), and in concert with a specified hydrograph, employs the Delft3D hydraulic model (Deltares, Netherlands; run here in two-dimensional mode for computational efficiency) to route water through the model domain and produce rasters describing flow depth, velocity, and bed shear stress. Sediment is entrained in cells where bed shear stress \( \tau_B \) exceeds the critical bed shear stress \( \tau_{BC} \) for the grain size in that cell (Buffington and Montgomery, 1997; Snyder et al., 2009):

\[
\tau_B > \tau_{BC}
\]

(4.1)

where the critical bed shear stress is computed as a function of relative sediment density \( \gamma \), acceleration due to gravity \( g \), and median particle size \( D \), and the Shields stress at incipient motion \( \tau_{\ast C} \), set equal to 0.05 such that:

\[
\tau_{BC} = \tau_{\ast C} \gamma g D
\]

(4.2)

The simulations described herein employed a constant grain size, set equal to 0.05 m \( (D_{50} \) for the study reach; Hodge et al., 2009). At each cell, sediment is scoured to a depth computed using the event-scale equation of Montgomery et al., (1996) that predicts scour depth as a function of bedload transport rate \( Q_b \), bedload velocity \( u_b \), sediment density \( \rho_S \); 2650 kg/m\(^3\)), and bed sediment porosity \( \gamma \):

\[
d_s = \frac{Q_b}{u_b\rho_s(1 - \gamma)}
\]

(4.3)

Once sediment is entrained, MoRPHED computes its transport downstream along flowpaths delineated using Delft3D-derived velocity vectors. At each cell along the flowpath, an amount of sediment is deposited as specified by a user-input path-length distribution (Figure 4.1), which describes the characteristic transport distances of entrained particles during a competent flow (Pyrce and Ashmore, 2003a,b; Kasprak et al., 2015b). Here we used a single-peak path length distribution with a total length of 210 m where the moment statistics \( \sigma \) and \( \mu \) were set to 125 and 40 respectively. Characteristics of the path length
distribution were obtained by measuring the average diffluence-confluence spacing across the study reach in the 2003 DEM, which has been shown to approximate the average travel distance of particles in braided streams (Pyrce and Ashmore, 2003b; Kasprak et al., 2015b). MoRPHED also includes routines for lateral bank erosion that scale bank retreat distance as a user-calibrated function of near-bank shear stress and bank slope, and these routines are further discussed in Kasprak et al. (2015a). MoRPHED outputs an updated DEM following each flood (Figure 4.1).

Because the concept of the sediment path length is inherently linked to individual competent flows (i.e. floods), the time step of model operation is equal to one flood, regardless of its duration. We modeled all competent flows in the study reach during the period 2003-2013; that is, hydraulics and sediment transport were computed for the 185 individual flood peaks in excess of 20 m$^3$/s at the study reach during this eleven-year period (Figure 4.2).

Sediment import is modeled in MoRPHED by employing the same path-length distribution as specified for bed sediment transport and deposition, beginning at the upstream reach boundary. Sediment can be imported up to once per model timestep (i.e. once per flood). The amount of sediment to import can be specified either as a proportion of exported sediment or as an absolute volume. Here we specified four end-member cases of sediment import: (a) upstream sediment supply set equal to exported sediment for each flood (e.g., sediment equilibrium), (b) upstream sediment supply set equal to one half of exported sediment volume (e.g., sediment deficit), (c) upstream sediment supply set equal to twice the volume of exported sediment (e.g., sediment surplus) and (d) upstream sediment supply set equal to exported sediment, but import occurred only once every ten events (e.g., periodic sediment import). Each of these four scenarios is detailed in Table 4.1.

4.4 Analysis Techniques

The output DEMs from each model scenario (Section 4.6, Table 4.1) were differenced to create DEMs-of-Difference (DoDs) at 3, 5, and 10-year intervals (Figures 4.4 -
Figure 4.2. River Feshie Study Reach and Discharge Record. Hillshaded DEM from 2003 is shown atop 1 m aerial imagery. Bottom panel shows downscaled 15-minute discharge record at Glenfeshie for the period 2003-2013. Estimated competent flow (20 m³/s) in the Glenfeshie reach shown in green.

4.7) with the 2003 Glenfeshie DEM using the Geomorphic Change Detection 6 software (http://gcd.joewheaton.org). Complete methodology for DEM differencing and geomorphic change detection are detailed in Wheaton et al. (2010) and Kasprak et al. (2015b). All changes in the 2003-2013 DoDs greater than a threshold limit of 0.2 m were manually
classified into one of eleven braiding mechanisms, which are shown in Figure 4.8 and detailed in Wheaton et al. (2013) and Kasprak et al. (2015b). Braiding mechanisms are the morphodynamic processes that act to create or maintain a braided planform, and may vary in response to shifting hydrologic regime or sediment delivery at timescales as short as an individual flood (Ashmore, 1991). The relative contribution of each braiding mechanism to total volumetric change was recorded, along with the overall reach-scale sediment budget

<table>
<thead>
<tr>
<th>Scenario</th>
<th>Name</th>
<th>Description</th>
<th>Applicable Case Studies in Braided Rivers</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Equilibrium</td>
<td>Sediment import at upstream reach boundary set equal to exported sediment volume</td>
<td>Numerous laboratory flume studies (Leopold and Wolman, 1957; Ashmore, 1991; Pyrce and Ashmore, 2003b; Kasprak et al., 2015b)</td>
</tr>
<tr>
<td>2</td>
<td>Deficit</td>
<td>Sediment import at upstream reach boundary set to half of exported volume</td>
<td>Dams or channel works, torrent control; (Roux et al., 1989; Gilvear, 2004; Erwin et al., 2011)</td>
</tr>
<tr>
<td>3</td>
<td>Surplus</td>
<td>Sediment import at upstream reach boundary set equal to twice the volume of exported sediment; sediment surplus</td>
<td>Climate-induced sediment delivery (Vandenberghé, 2002; Vandenberghé, 2003); land-use change (Wyzga, 1993)</td>
</tr>
<tr>
<td>4</td>
<td>Periodic Import</td>
<td>Sediment import at upstream reach boundary set equal to exported sediment, but import only occurs once every ten floods; sediment equilibrium</td>
<td>Sediment sourced from rare tributary floods (Erwin et al., 2011); gravel mining within a reach (Piégay et al., 1997); conditions of net deficit over intermediate periods (Wheaton et al., 2013)</td>
</tr>
</tbody>
</table>
4.5 Results

Results of decadal-scale geomorphic change and modeling using each of the four scenarios in Table 4.1 are shown in Figure 4.3. The progression of geomorphic changes in model runs at 3, 5, and 10-year intervals are shown in Figures 4.4 - 4.7. In field observations, along with the equilibrium and periodic model simulations, a multi-thread channel planform was maintained; the deficit scenario produced a system with a single dominant anabranch, whereas the surplus scenario produced widespread deposition and loss of distinct channels across the model domain. Field-based DEM differencing (Figure 4.3A) revealed geomorphic change distributed across the braidplain, and generally marked by low-magnitude elevation changes, typically confined to 1.5 m of erosion or deposition. Mechanistic segregation of the field-based DoD (Figure 4.8A) indicates that a diversity of braiding mechanisms were responsible for geomorphic change in the study reach; the combination of central bar development (26%), transverse bar conversion (17%), lobe dissection (16%), and bank erosion (16%) were responsible for the majority of volumetric change through the study reach.
Figure 4.3. Sediment Supply Scenario Modeling Results. DEM differencing from field measurement during 2003-2013 period (A) is compared with results of scenario-based modeling for each sediment supply scenario (B-E) using reach-scale mapping and elevation change distributions.
Figure 4.4. Results: Equilibrium Simulation. DEMs-of-Difference (top row) and elevation change distributions (bottom row) for equilibrium scenario at 3, 5, and 10 year intervals.
Figure 4.5. Results: Deficit Simulation. DEMs-of-Difference (top row) and elevation change distributions (bottom row) for deficit scenario at 3, 5, and 10 year intervals.
Figure 4.6. Results: Surplus Simulation. DEMs-of-Difference (top row) and elevation change distributions (bottom row) for surplus scenario at 3, 5, and 10 year intervals.
Figure 4.7. Results: Periodic Simulation. DEMs-of-Difference (top row) and elevation change distributions (bottom row) for periodic scenario at 3, 5, and 10 year intervals.
Results of scenario-based morphodynamic modeling varied widely depending on the scenario being simulated. In general, depth of deposition across the model reach agreed well with field-observed values; maximum aggradation depth averaged 2.3 m across the four scenarios, compared to 2.1 m via field observations. The form of depositional change, as detailed via elevation change distributions (Figure 4.3B-4.3E), also mirrored that seen in the field, with the majority of depositional changes occurring in thin-mantled areas across the braidplain. Geomorphic change due to erosion varied widely between the four scenarios. Erosional change was smallest in the surplus scenario (4,736 m$^3$ of total erosion) and was most pronounced in the deficit scenario (25,320 m$^3$ of total erosion). The form of erosional changes in the equilibrium and periodic scenarios mirrored those seen in the field (elevation change distributions in Figure 4.3), but were generally greater in volume than erosional changes observed via field-based DEM differencing (average 17,967 m$^3$ in models, 12,680 m$^3$ in field). In these runs, the volume of changes due to deposition was comparable to that seen in the field (average 12,770 m$^3$ in models, 14,677 m$^3$ in field).

Generally, a smaller number of braiding mechanisms were responsible for a greater amount of geomorphic change in the model runs than in the field. This is particularly true for the deficit and surplus runs, where channel incision (76% of change) and transverse bar conversion (45%) were responsible for a large proportion of geomorphic change respectively. The equilibrium and periodic scenarios produced a greater diversity of braiding mechanisms, and those mechanisms were generally responsible for smaller volumes of geomorphic change. In these scenarios, the contribution of individual braiding mechanisms was generally similar to that seen in the field, although notable differences were present. In particular, bank erosion played a reduced role in scenario-based modeling as compared to field-observed changes (17% of total change in the field compared to average 2% in models). Conversely, channel incision was more pronounced in modeling than was observed in the field (4% in field, average 23% in models).
Figure 4.8. Mechanistic Segregation of Decadal Modeling Results. Delineated braiding mechanisms and their relative volumetric contributions to total geomorphic change are shown for field observations (A) and each of the four sediment supply scenarios (B-E).
4.6 Discussion

The geomorphic change produced during scenario-based model runs is intuitive given the degree to which the study reach is shifted into sediment surplus or deficit over the course of the simulation. When shifted into sediment surplus or deficit, a transition away from the braided planform was observed (via overbank deposition or channel incision, respectively). Alternatively, when sediment equilibrium was maintained, so too was the braided planform. Mueller and Pitlick (2014) likewise found that braided and single-thread reaches could be readily separated based on the amount of sediment transported through reaches during floods; using a discriminant function, they were able to predict channel planform in 51 of 53 study reaches, highlighting the relative importance of sediment supply. Our modeling results demonstrate that relatively small variations in sediment supply can have profound effects on resultant channel planform: in our scenario-based simulations, a twofold variation in the amount of supplied sediment was sufficient to produce large-scale channel pattern variability over a period of ten years. Specifically, the surplus run produced widespread deposition of material over the braidplain, resulting in a reach marked by large swaths deposited material and an overall loss of the original channel network (Figure 4.3C). In contrast, the deficit run transformed the reach from a multi-threaded network to one marked by a single main channel that underwent a large degree of incision over the course of the model run (Figure 4.3D), reminiscent of braided-to-single-thread transitions of channels that have seen reductions in sediment supply (Gurnell et al., 2009). Between these two end-member scenarios are the equilibrium and periodic model scenarios, which, despite preferential incision of the two main anabranches, maintained the initial multi-threaded character of the reach (Figures 4.3B, 4.3E); the magnitudes and form of erosion/deposition in these two scenarios most closely mirrored those seen in the field during the same ten-year period (Figure 4.3A).

Simply put, the results of scenario-based modeling on the Feshie indicate that sediment supply is extremely important in determining the form of a reach, maintaining a multi-threaded planform, and in Kasprak et al. (2015a) explored the relative influence of varying
the path length distribution employed in MoRPHED model runs at annual scales, and found that in general, the choice of distribution exerts relatively little influence on the overall geomorphic change across a model reach; this stands in contrast to the high variability and rapid divergence between simulations here depending on the sediment supply scenario employed.

Sediment supply becomes manifest in planform changes by altering the relative contributions of individual braiding mechanisms that drive erosion and deposition. For example, in the segregated DoD obtained from field observation (Figure 4.8A), all eleven possible braiding mechanisms were observed, with no individual process accounting for more than 26% of volumetric change in the study reach. Similarly, in the equilibrium and periodic scenarios, the greatest percent contribution of any braiding mechanism was 26% (channel incision). On the other hand, in the deficit and surplus scenario simulations, where channel incision (76%) and transverse bar conversion (46%) respectively comprise large proportions of the volumetric change across the model reach.

MoRPHED is a model, with shortcomings of course; in particular, the model appears to under-predict the amount of bank erosion (average 2 in model runs, 15% in field) and over-predicts the amount of channel incision relative to field observations (average 31% in model runs, 4% in field). Despite this, the model reproduces the complete diversity of geomorphic processes (e.g., braiding mechanisms) that lead to the development and maintenance of multi-thread channels in the field. Because of this general though imperfect reproduction of form and process (Paola et al., 2009), MoRPHED provides an opportunity to better understand the importance of sediment source on the maintenance of braided channels. To this effect, it appears that maintenance of the braided planform is dependent on sediment sourced both locally (e.g., within the reach) and from upstream.

That braiding can be maintained through reworking of locally-sourced sediment is evidenced by the fact that the multi-threaded planform persists despite an absence of upstream supply, particularly in the periodic simulation, and over short periods (e.g., 3 years; Figure 4.5) in conditions of sediment deficit. Indeed, certain braiding mechanisms that scour large
volumes of material from within a reach (e.g., bank erosion, bar edge trimming, channel incision) may provide sediment for bar building within that reach provided that sediment travel distances do not result in eroded sediment being immediately exported from the reach. In the case of these model scenarios, the mean travel distance of 125 m, combined with the reach length of \( \sim 650 \text{ m} \), imply that on average, particles eroded at the upstream end of the model reach would remain within the reach for roughly five events, providing opportunities for eroded sediment to be deposited within the model domain. In some instances, sediment sources and sinks can be directly correlated, given that we know the mean particle travel distance. Particularly in the equilibrium and periodic model runs, areas of sediment accumulation marked by transverse bar conversion and central bar development were located directly downstream from sediment sources marked by lobe dissection and channel incision.

Despite the importance of local sediment sources in the maintenance of the braided planform, these sources cannot overcome extended periods of sediment deficit. In our simulations, an initially multi-thread planform was reduced to a reach marked by a single main anabranch within ten years (185 floods in excess of 20 \( \text{m}^3/\text{s} \)) in conditions of sediment deficit. Thus, a supply of sediment delivered from upstream that roughly mirrors the volume of exported sediment appears vital in maintenance of the multi-thread planform. Interestingly, however, the planform appears relatively insensitive to the periodicity with which that sediment is delivered. Multi-thread channel planform was preserved in both the equilibrium and periodic simulations (Figure 4.3), and a diversity of braiding mechanisms were produced (Figure 4.8), despite sediment being imported every flood or every ten floods, respectively. The maximum interval with which sediment can be delivered to the reach before the multi-thread planform begins to shift to a single-thread channel is deserving of further research; as a starting point, our results indicate that delivery once every ten floods (or roughly once per year on the Feshie) is sufficient for planform maintenance.

Finally, it is reasonable to consider whether the magnitude of channel change predicted by MoRPHED aligns to a reasonable degree with the timescale over which those changes occurred. In all simulations, several meters of erosion or deposition were observed, along with
large-scale channel planform transitions, particularly in the deficit and surplus simulations. Nominally, all of these geomorphic changes took place over a ten-year timespan. Drastic changes in channel planform, particularly arising from a sudden influx or loss of sediment on braided streams, are common over decadal periods (Gurnell et al., 2009). However, because MoRPHED is not directly linked to discrete units of time, but rather simulates evolution using an event-based model timestep, it is difficult to determine whether similar magnitudes of geomorphic change would take place on the Feshie under altered sediment supply scenarios as modeled herein. Some insight may be gained by comparing the characteristic volumes and depths of erosion and deposition in the relatively undisturbed model scenarios (e.g., equilibrium and periodic) to values from field observations. In general, MoRPHED produced volumes and characteristic depths of deposition that align reasonably well with field-observed values (Figure 4.3). The average depth of deposition was 0.53 m for the model (12,770 m$^3$ total), and 0.41 m in the field (14,677 m$^3$ total). On the contrary, MoRPHED consistently over-predicted the depth and volume of erosion as compared to field data (Figure 4.3). The average depth of erosion was 0.80 m for the model (17,967 m$^3$ total) and 0.53 m in the field (12,681 m$^3$ total), suggesting that geomorphic change due to erosion may be artificially accelerated in MoRPHED. This divergence in prediction validity is not surprising: a good deal of research has linked channel morphology and particle deposition locations, which form the basis for the depositional component of MoRPHED (Sear, 1996; Habersack, 2001; Pyrce and Ashmore, 2003a,b; Kasprak et al., 2015b). Our understanding of event-based sediment scour depth remains limited, as does the geomorphic community’s ability to predict depth of scour during a flood (Montgomery et al., 1996). To that end, our ability to scale event-based morphodynamics to discrete units of time remains hindered, as we cannot reliably estimate geomorphic change due to erosion on a per-flood basis. If simplified models such as MoRPHED are to be used deterministically, refinement of event-scale scour prediction is vital.
4.7 Conclusions

The volume of sediment supplied to a given reach is a reliable discriminator of channel planform (Mueller and Pitlick, 2014), and small variations in sediment supply can produce strikingly different channel planforms. This research employed the MoRPHED morphodynamic model to investigate channel planform change in response to varied sediment supply on the River Feshie, UK. Four scenarios were modeled and compared to field observations over a ten-year period, including sediment equilibrium, sediment deficit, sediment surplus, and periodic sediment import. Of these, surplus and deficit resulted in large-scale channel pattern transitions away from a multi-threaded planform. The equilibrium and periodic simulations maintained the braided character of the reach, despite significant variability in the timing of sediment delivery (each flood compared to every ten floods), while producing a strikingly similar suite of braiding mechanisms. This suggests that while sediment supply is vital in braided planform maintenance, a multi-thread system may persist (at least over decadal timescales) through reworking and bar building using sediment sourced from within the reach. The maximum timescale over which local sediment can maintain a multi-thread planform is deserving of further study. This research highlights the utility of simplified models as learning tools for fluvial dynamics, while also illustrating the need for an improved understanding of event-scale erosion prediction if such models are to be used in a more deterministic fashion in the future.

4.8 Acknowledgments

This research is supported by a National Science Foundation grant (No. 1147942, ‘Sensitivity of Braided River Morphodynamics to Sediment Supply’). Field data collection on the Feshie was completed with the help of James Brasington, Niall Lehane (Queen Mary, University of London), Rebecca Hodge (Durham University), Julian Leyland, Davide Aramini (Southampton University), Mark Smith (Leeds University), and Damia Vericat (Forest Technology Institute of Catalunya), with the generous permission of Thomas MacDowell and the Glenfeshie Estate. Sara Bangen, Eric Wall, Kenny DeMeurichy (Utah State
University), and Philip Bailey (North Arrow Research) assisted with DEM generation and differencing.
References


Braided rivers are highly dynamic systems, marked by multiple channels splitting and rejoining around a network of transient bars [Ashmore, 1991; Brierley and Fryirs, 2005]. They are characterized by high sediment availability, rapid bank erosion, and dynamic channel evolution. Understanding the potential response of braided rivers to altered sediment or hydrologic regimes is difficult because morphology may adjust across a range of timescales, often ranging from individual floods to centuries [Gurnell et al., 2009], meaning that traditional field observation may not capture channel response to water or sediment supply changes. Here I developed a numerical model for braided river evolution based on characteristic particle travel distances revealed via laboratory flume experiments. I subsequently employed this model to investigate the response of braided channels to altered sediment supply at event to decadal timescales.

5.1 Summary

In Chapter 2, I employed laboratory flume experiments to measure the travel distances of fluorescent tracer particles over the course of five simulated floods on a braided channel. I documented channel evolution during floods using digital elevation models derived from structure-from-motion photogrammetry. I sought to understand whether there was a relationship between channel morphologic units, specifically the location of channel bars, and the deposition locations of tracer particles, as hypothesized by prior researchers [Pyrce and Ashmore, 2003a, 2003b]. The results of these experiments revealed a close coupling between the location of bars, specifically the heads and margins of mid-channel and lateral bars, with the deposition locations of tracer particles: 81% of recovered tracers were found in association with bars, and of these, 70% were deposited on bar heads and margins. Moreover, the mean particle travel distance of 2.5 m was closely related to the average confluence-diffluence spacing (2.3 m) in all runs, suggesting an intrinsic coupling between
channel morphology and event-scale particle transport distance, or path length, in braided rivers.

In Chapter 3, I developed a morphodynamic model for gravel bed braided rivers that leverages the relationships between path length and channel morphology revealed in Chapter 2. The model couples the Delft3D hydrodynamic model with a novel sediment transport/morphodynamics algorithm in order to simulate channel evolution at the scale of individual floods. I verified the model’s performance using a plurality of approaches, including digital elevation model differencing, segregation of braiding mechanisms, and channel morphometric indices compared to field-measured values in two braided rivers [e.g., Thomas and Nicholas, 2002; Lesser et al., 2004; Nicholas, 2013]. The developed model demonstrated the ability to produce all of the braiding mechanisms observed in the field, while also reproducing the general form of elevation change distributions observed via repeat topographic surveys. At the same time, the magnitudes of change predicted by the model often differed from field observed values (e.g., exaggerated change on the Feshie, muted change on the Rees), and the relative contribution of individual braiding mechanisms did not always accurately reflect those observed in the field. The results of this chapter indicate that despite being necessarily imperfect, simplified models of braided river morphodynamics can lend insight with regard to the suitability of various process representations (e.g., modeling frequency, choice of path length distribution) for predicting channel evolution. Further, the results indicate the necessity of developing improved predictions of event-scale scour depth and lateral migration if models that operate at the timescale of individual floods are to more accurately reproduce channel evolution observed in the field.

In Chapter 4, I employed the model developed in Chapter 3 to examine the morphodynamic response of a braided river to variations in sediment supply from upstream. I modeled channel evolution over a ten-year period under four sediment supply scenarios: equilibrium, deficit, surplus, and periodic import occurring once every ten floods. Of the four scenarios that were modeled, only the equilibrium and periodic import model simulations maintained a braided planform after the ten year period; the deficit simulation resulted in a deeply-
incised single-thread channel planform, and the surplus scenario produced widespread sheet deposition over the braidplain and loss of a distinct channel network. The results of this modeling suggest that in general, braided planform can be maintained under conditions of sediment equilibrium. At the same time, channels do not appear to be highly sensitive to the frequency with which sediment is imported. Over decadal timescales, local supplies of sediment (stored in channel bars) may counteract a lack of supply from upstream and allow for persistence of the braided planform, suggesting that braided channels may be inherently buffered against alterations in sediment supply. However, if supply remains limited over decadal timescales, alterations to the braided planform and a general shift toward a single-thread channel may occur.

5.2 Synthesis

5.2.1 Linking Channel Morphology and Particle Travel Distance

This research was motivated by a desire to more fully elucidate relationships between sediment transport and channel morphology in braided rivers, and to determine whether those relationships could inform simple morphodynamic models. The laboratory flume experiments in Chapter 2 revealed the close coupling between the location of in-channel geomorphic units, particularly channel bars, and the deposition locations of particles in transport. Concurrently, the experiments revealed that particle travel distance could be approximated using simple morphometric parameters, namely the spacing between diffluence-confluence couplets in a braided reach.

The laboratory experiments conducted as part of Chapter 2 build on previous research documenting particle path length relationships with geomorphic unit spacing in braided and single thread channels [Sear, 1996; Habersack, 2001; Pyrce and Ashmore, 2003a, 2003b]. In so doing, the research resulted in a conceptual framework for predicting downstream transport distances as a function of confluence-diffluence spacing, which is deserving of further investigation to determine its suitability across a variety of braided systems. To that end, it is important to consider the degree to which these laboratory experiments
accurately reflected transport conditions in the field. The flume at the University of Western Ontario is a Froude-scale model of the gravel-bed Sunwapta River [Hundey and Ashmore, 2009], and Froude modeling is a robust tool for replicating flow hydraulics and sediment transport in actual systems [Warburton, 1996]. At the same time, three components of the experiments should be further investigated in future research to determine the nature of the path length - channel morphology relationship across a range of discharges, sediment sizes, and flood duration. First, because of the limited time available for laboratory experiments, a constant discharge was used in all flume runs; results indicate a close coupling between particle travel distance and in-channel geomorphic units at this ‘channel-forming’ discharge, but the degree to which this relationship holds across a range of competent flows is unclear. Second, each of the flume runs lasted for 20 minutes in duration, and while this run time was sufficient for downstream mobilization of tracer particles, it did not result in the migration of in-channel bars nor dramatic reworking of the braided network. Given this minimal channel evolution during flume runs, it is unclear whether the coupling between transport distance and geomorphic unit spacing will persist as geomorphic units evolve over more extended floods. Finally, the tracer particles employed in Chapter 2 were representative of the coarser component of the bed material, approximated 2D_{50}. While transport distances may be relatively insensitive to particle size up to these diameters [Church and Hassan, 1992], experimentation using a wider variety of particle sizes is merited to determine whether deposition location and transport distances are similar across the range of particle diameters seen in field settings. This is particularly true for the fine end-member of the particle size distribution in natural channels, which were omitted from the flume experiments performed here to avoid the use of silt and finer particles in the laboratory.

5.2.2 Imperfect Models as Learning Tools

Despite the necessarily simplified nature of the laboratory experiments, the relationship between particle travel distance and geomorphic unit spacing that resulted from the research in Chapter 2 laid the foundation for an efficient morphodynamic model in Chapter
3. This model was subsequently employed for scenario-based investigation of the influence of sediment supply on reach-scale decadal morphodynamics in Chapter 4. The model, which computes channel evolution by transporting sediment according to user-specified path-length distributions in concert with hydraulics based in computational fluid dynamics (CFD), is the first of its kind. This 'hybrid' approach that includes aspects of cellular reduced complexity models [e.g., Murray and Paola, 1994] and more intensive CFD-based routines [Bates et al., 2005] is unique in that it attempts to leverage the most advantageous components of each strategy, resolving bar-scale morphodynamics at decadal timescales.

Comparing the results of morphodynamic modeling to channel evolution observed via repeat field surveys reveals that this highly simplified approach is capable of reproducing the volumetric signatures of deposition, along with field-observed braiding mechanisms. At the same time, the volumes and spatial patterns of change diverge significantly from those seen in the field, particularly over more extended (e.g., decadal) timescales. The model is imperfect, as are all models in the earth sciences [Oreskes et al., 1994]. However, this imperfection is to be expected given the highly-simplified nature of the code; to simply focus on the instances when the model is incorrect is to ignore its potential to provide insight with regard to (a) formative processes in braided rivers and (b) the need to parameterize certain processes in morphodynamic modeling versus those processes that result without explicit algorithmic representation. Additionally, it may be unreasonable to expect exact planform reproduction in this simplified morphodynamic model over annual and longer timescales, and instead comparisons such as mechanistic segregation or morphometric analyses may be more appropriate for validation (see also Chapter 3).

With regard to the former, decadal scale modeling on the River Feshie (Chapter 4) revealed drastic planform changes resulting from variation in the volume of upstream sediment supplied to a braided reach. While I do not expect the individual model outputs to perfectly replicate those conditions that would be observed in the field under these sediment supply scenarios, the planform changes do mirror those seen in braided rivers following a shift in supply regime (e.g., transition to single-thread channel following reduction in supply;
Perhaps more importantly, the finding that braided planform maintenance can be achieved under both strict equilibrium (import equals export for each flood) and equilibrium under periodic import conditions points to the importance of reach-scale sediment supply in maintaining a braided network.

With regard to emergent versus parameterized process representation, the development of a novel modeling routine such as that in Chapter 3 led to uncertainty regarding which channel behaviors (e.g., braiding mechanisms) would emerge simply as a result of the flow field and boundary stresses versus which mechanisms would need to be explicitly incorporated into the model algorithm (e.g., parameterized). The results of this research indicate that apart from bar edge trimming and bank erosion, each of the field-observed braiding mechanisms occurred in morphodynamic modeling simply as a result of interaction between the flow field and channel bed. Because bar edge trimming and bank erosion result from scour of surfaces that may not be inundated (that is, \( \tau_B = 0 \)), they require explicit parameterization in the form of a lateral retreat algorithm. Finally, the fact that all braiding mechanisms were reproduced via modeling does not mean that the model accurately reflected their individual volumetric contribution to total change. This is especially true in the case of bank erosion and chute cutoff on the Feshie (Chapters 3 and 4), which were underestimated when compared to field-derived mechanistic segregation. As a result, setting-specific calibration of certain model components, or explicit incorporation of a model algorithm increasing chute cutoff, may be required to more closely mirror morphodynamic evolution in the field. Nevertheless, the ability of this highly simplified hybrid modeling approach to replicate braiding mechanisms observed in the field is promising.

5.2.3 Ways Forward in Simplified Morphodynamic Modeling

This research represents a first step in the development of morphodynamic models that explicitly couple sediment transport and channel morphology, and in so doing, paves the way for computationally-efficient algorithms at spatiotemporal scales that match those of field observation [Hafen et al., In Prep.]. At the same time, the model would benefit from a
more refined understanding of (a) event-scale scour depth and (b) lateral channel migration, and their incorporation into numerical models, within the geomorphic community.

In the case of event-scale scour depth, the morphodynamic model developed herein employs a highly simplified routine for computing the volume of scour on a cell-by-cell basis [Montgomery et al., 1996]. While estimation of scour depth during a competent flow is notoriously difficult, accurately predicting the amount of material removed across the model domain has major implications for the validity of the morphodynamic approach developed here. It is striking that in nearly all simulations, the developed model is able to reproduce the form of volumetric changes resulting from deposition (e.g., elevation change distributions in Chapters 3 and 4). Divergence between model results and field observations primarily manifests in the erosional component of change, and in the overall volume of geomorphic change during a simulation. In the case of the latter, this is presumably a direct effect of inaccuracy in computing the volume of scour, as the volume of deposition is equal to the volume of erosion in the model (e.g., conservation of sediment mass). As a result, the refinement of the developed morphodynamic model from a simplified learning tool to a deterministic utility for particular field settings would greatly benefit from an improved understanding of event-scale bed scour within the geomorphic community.

In the case of lateral channel migration, the propensity for braided rivers to readily adjust their boundaries (e.g., bank and bar erosion) is a hallmark of the planform [Ashmore, 1991]. At the same time, deterministic modeling of channel migration is notoriously difficult [Rinaldi and Darby, 2007]. Whereas small-scale approaches that compute the balance between the resistance of banks to downslope failure (e.g., a factor-of-safety approach) are available, methods for the prediction of bank retreat at the reach scale, particularly over the course of a competent flow event, are largely unavailable [see review in Rinaldi and Darby, 2007]. The model developed herein contains routines for bank retreat, although they must be empirically calibrated depending on the site being modeled. Even with empirical calibration, bank retreat and bar edge trimming are underpredicted in our morphodynamic model. As a result, the utility of this framework would greatly benefit from an improved ability to
predict lateral channel migration distances at the reach and event scale. It is possible that the inclusion of a subgrid routine for bank and bar retreat would allow for bank migration at distances less than the resolution of the model DEM, although these approaches are also in their infancy [e.g., Ganti et al., 2012] and require further refinement for inclusion into this modeling framework.

Finally, the morphodynamic model used here was limited to simulations involving a single, reach-scale representative grain size. Although the mathematics underlying prediction of sediment entrainment should remain constant across a range of gravel-sized sediment [e.g., Buffington and Montgomery, 1997], our use of a single grain size had implications for the development of certain stratigraphies common to braided rivers. These include development of a coarse armor layer on the bed of the stream [Dietrich et al., 1989; Guerit et al., 2014] that increases critical shear stress for particle entrainment and may decrease scour depth during some events. Additionally, the development of composite banks that may exhibit variable resistance to mass failure based on their stratigraphy [Dapporto et al., 2003; Brierley and Fryirs, 2005] was not possible using this simplified modeling routine. Finally, our model does not account for the presence of fine sediment (e.g., sand and finer fractions) at all. Given that fines can alter transport rates of gravel [Wilcock et al., 2009], future simulations would benefit from the inclusion of a fine sediment component. At the same time, the degree to which fine sediment path lengths can be characterized by channel morphology, or whether fine sediment is transported downstream with characteristic distances at all, remains unclear and is deserving of further investigation.

5.3 Conclusion

This research demonstrates that simplified relationships between particle travel distance and channel morphology in braided rivers can be integrated into morphodynamic models. Further, the modeling framework developed here is a learning tool that, despite imperfection, provides insight with regard to the explicit parameterization of processes in morphodynamic modeling, the importance of upstream sediment supply in planform main-
tenance, and the potential for reach-scale sediment supply to provide resilience against braided planform loss in light of sediment supply alteration. The modeling framework developed here is the first of its kind in that it fuses reduced complexity and computational fluid dynamics approaches to model morphodynamics at the event scale. This approach, limited here to the reach scale in gravel-bed rivers, has the potential to allow for efficient prediction of morphodynamic evolution at spatiotemporal scales consistent with field observations of channel change. It is intended to provide a first step in a new approach for morphodynamic modeling of river systems, and with further research into event-scale scour depth and channel migration, has the potential to be a valuable site-specific, deterministic modeling utility for predicting channel evolution.
References


Hafen, K., A. Kasprak, and J.M. Wheaton (In Preparation), Methods for modeling morphodynamics at the event scale to increase computational efficiency.


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Alan Kasprak
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Thanks,

Alan

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James Hensleigh
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Dear James,

I am in the process of preparing my dissertation in the Watershed Sciences Department at Utah State University. I am requesting your permission to include the manuscript we co-authored, *The relationship between particle travel distance and channel morphology: results from physical models of braided rivers*, which was published in *Journal of Geophysical Research: Earth Surface* in 2015.

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APPENDIX C Permission-to-Use Letters, Chapters 3 and 4
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Konrad Hafen
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Dear James,

I am in the process of preparing my dissertation in the Watershed Sciences Department at Utah State University. I am requesting your permission to include two manuscripts we coauthored, *Coming to grips with model imperfection: morphodynamic models as exploratory tools for understanding braided river dynamics* and *The sensitivity of braided river morphodynamics to variations in sediment supply and source*, which will be submitted for publication in 2015.

Please indicate your approval of this request by signing in the space provided below. If you have any questions, please don’t hesitate to contact me.

Thanks,

Alan

Alan Kasprak
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I hereby give permission to Alan Kasprak to reprint the following material in his dissertation:

Kasprak A, Brasington J, Hafen K, Wheaton JM. In Preparation. *Coming to grips with model imperfection: morphodynamic models as exploratory tools for understanding braided river dynamics*.


Signed ________________________________________________  Date _________________

1st September 2015

Dr. Brasington

1st September 2015
VITA

Alan Kasprak

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EDUCATION

Doctor of Philosophy in Watershed Sciences 2015
Emphasis in Geomorphology and Earth Surface Processes
Utah State University - Logan, UT

Master of Science in Earth Sciences June 2010
Dartmouth College - Hanover, NH
MS Thesis: Stream Channel and Riparian Response to Land-Use in Northern New England

Bachelor of Science in Geology/Geophysics May 2008
Boston College - Chestnut Hill, MA
BS Thesis: Measuring Sedimentation Rates and Land-Use Change in a Dam-Influenced Lake Delta: Narraguagus River, Maine

TEACHING AND RESEARCH EXPERIENCE

Utah State University - Logan, UT 9/10 - Present
Research Assistant

Dartmouth College - Hanover, NH 7/08 - 7/10
Teaching & Research Assistant

Boston College - Chestnut Hill, MA 5/07 - 8/08
Fluvial Systems Lab & Field Researcher

PEER-REVIEWED PUBLICATIONS


PEER-REVIEWED PUBLICATIONS - CONTINUED


PEER-REVIEWED PUBLICATIONS IN PROGRESS


SCIENTIFIC REPORTS

MEETING ABSTRACTS - PRIMARY AUTHOR ONLY


COURSES TAUGHT

CO-INSTRUCTOR

Intermountain Center for River Restoration and Rehabilitation
   Geomorphic Change Detection: Restoration Monitoring 2011 & 2014

Utah State University Watershed Sciences Graduate Induction Course
   An Introduction to Stream and Landscape Classification 2014

TEACHING ASSISTANT

Utah State University - Watershed Sciences Department
   Watershed Sciences Graduate Induction Course 2012 & 2013

Intermountain Center for River Restoration and Rehabilitation
   Geomorphology and Sediment Transport in Channel Design 2011

Dartmouth College - Earth Sciences Department
   Introduction to Earth Science 2008 & 2010
   Off-Campus Program (Western US Geology) 2009
   Oceanography 2009
   Earth’s Past, Present, and Future Climate 2009

LABORATORY TEACHING ASSISTANT

Dartmouth College
   Introduction to Earth Science Laboratory 2008 & 2010

GUEST LECTURES AND SEMINARS

USGS Grand Canyon Monitoring and Research Center 2015
   Linking Sediment Transport and Channel Morphology in Braided Rivers

Utah State University, Fluvial Hydraulics and Ecohydraulics Graduate Course 2014
   Introduction to Two-Dimensional Eco-Hydraulic Modeling

Utah State University, EcoLunch Brown Bag Seminar 2012
   Life, Landscape, and the Dynamic Nature of Physical Habitat

Dartmouth College, Off-Campus Program (Western U.S. Geology) 2009
   Sediment transport in the Grand Canyon

Dartmouth College, Off-Campus Program (Western U.S. Geology) 2009
   Ephemeral stream morphology in Death Valley

Dartmouth College, Geolunch Brown Bag Series 2009
   Anthropogenically-Driven Fluvial Geomorphology

GRANTS, AWARDS, HONORS

Utah State University
   Doctoral Dissertation Completion Award Grant Recipient ($20,000)

National Science Foundation
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   Co-Authored with PI JM Wheaton Grant Recipient ($271,000)

10th Federal Interagency Sedimentation Conference, 2015
   Best Student Technical Paper Award Winner

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   Graduate Student Travel Grant Grant Recipient ($300)
GRANTS, AWARDS, HONORS - CONTINUED

Society for Sedimentary Geology
Graduate Student Travel Grant Grant Recipient ($500)

Geological Society of America
Graduate Student Research Grant Grant Recipient ($1000)

Dartmouth College, Office of Graduate Studies
Graduate Student Research Grant Grant Recipient ($2500)

Geological Society of America
Graduate Student Travel Grant Grant Recipient ($500)

Dartmouth College, Office of Graduate Studies
Presentation Travel Grant Grant Recipient ($300)

Boston College Geology & Geophysics Department
Best Undergraduate Research Presentation Award Winner

Geological Society of America, Northeastern Section
Undergraduate Travel Grant Grant Recipient ($100)

CONFERENCE SYMPOSIA CONVENED
