ABSTRACT

LYONS, NATHAN JAY. Hillslope-Stream Coupling in Tectonically Active and Inactive Regions. (Under the direction of Karl W. Wegmann).

Climate, tectonics, and anthropogenic activity drive topographic change in a landscape. Stream and hillslope processes, which are modulated by these drivers, conduct much of the change. Significant opportunities remain to advance conceptual understanding of how shifts in the magnitude and character of these drivers is transmitted between streams and hillslopes, including how spatial gradients of precipitation and tectonic uplift affect patterns of basin-wide erosion, which processes drive erosion where streams and hillslopes are decoupled due to changes in land use, and what is the relative share of lithology and fluvial sediment transport in controlling sediment grain size distribution along the lengths of streams.

The processes acting across hillslope and channel domains of three settings were investigated. In the first chapter, hillslope and stream coupling in response to millennium-scale stream incision was assessed in the Clearwater River basin, Olympic Mountains, Washington State. A numerical model that quantified the positions of focused stream incision was constructed using topographic data inputs and temporal constraints provided by previous geologic studies. Stream incision into a late Pleistocene terrace initiated a wave of erosion that is now expressed as steepened slopes on hillsides, and as knickpoints on streams.

In the second chapter, the combined hydrometeorological conditions and dominant processes driving streambank erosion were investigated in the Atlantic Piedmont region of the eastern United States. A streambank was analyzed using data from
stream stage, precipitation, and streambank face topography over 19 months between 2010 and 2012. In both the monitoring period and field-based experiment, the largest volume of erosion occurred where sediment columns detached along vertical desiccation and horizontal seepage cracks. The results of this study demonstrated that anthropogenic sediment produced in historic times alters both the volume of fine sediment that enters streams and the processes that dominate streambank erosion.

In the final chapter, the longitudinal distribution of streambed sediment grain size was investigated in the Sandhills region of the Atlantic Piedmont. This study is motivated by the decline of numerous freshwater mussel species in the southeastern United States. Freshwater mussels prefer stable streambeds, which is largely controlled by the sizes of grains on the streambed. The longitudinal grain size distribution in three streams was modeled using theoretical and empirical sediment transport relationships with primarily topographic data inputs. Streams in this region are strongly dominated by sand or gravel. For this reason, the model attempted to predict median streambed grain size in two grain size fractions: sand (<2 mm) and gravel (>2 mm). Sediment samples were collected along the stream to evaluate this model, which correctly predicted all but one of the samples using this two fraction approach. This model can benefit stream restoration efforts and investigations of aquatic habitat availability.
Hillslope-Stream Coupling in Tectonically Active and Inactive Regions

by
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DEDICATION

This dissertation is dedicated to my parents, Ilene Wardle and Jay Lyons. They have encouraged me to follow my interests, believe in my abilities, and to be curious.
BIOGRAPHY

I was born in south Florida, and upon witnessing topographic variation and complexity in North Carolina, pursued a bachelor’s degree in geology at Guilford College. A position with the National Park Service during the summer prior to his first semester of graduate studies led to a Masters degree at North Carolina State University. My interest in fluvial, hillslope, and tectonic geomorphology grew during this time, which drove me to pursue a Ph.D. in a range of topics within this field.
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Numerous people and organizations provided encouragement, advice, and support during the production of this dissertation. Here, I acknowledge many of these people.

Jacqueline Gronwald, my wonderful girlfriend, was incredibly supportive throughout my graduate studies. The constant love and encouragement that Jackie provided were vital to the completion of this dissertation. She even pitched in during field work.

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Karl Wegmann, the chair of my dissertation committee and academic advisor, is incredibly devoted to his roles as an educator, advisor and scientist. My graduate education greatly benefitted from this devotion, and his passion for and knowledge of the geosciences.
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Chapter 1

Deconvolving Climatic and Tectonic Components of Transient Erosion Signals in the Olympic Mountains

1.1. Abstract

Stream power models have been applied in multiple settings often with assumptions of uniform tectonic uplift and precipitation. Recent investigations have explored the influence of precipitation gradients upon river longitudinal form using these models. Important opportunities for conceptual advancement remain in regions where both uplift and precipitation vary throughout a basin. Using numerical modeling and temporal constraints provided by previous geologic studies, we assess hillslope and stream coupling in response to millennium-scale stream incision in the Clearwater River basin, Olympic Mountains, Washington State. A model of knickpoint propagation indicates that incision into a late Pleistocene terrace initiated a wave of erosion that is now expressed as steepened slopes on hillsides, and as knickpoints on streams. Knickpoint vertical velocities are between 0.07–1.0 mm/yr, which is similar to previously reported rates of stream incision (0.01–0.8 mm/yr) and exhumation (0.03–0.8 mm/yr) in this basin. A chronometer independent from stream incision, exhumation, and knickpoint propagation rates was established using the genetic divergence of Cutthroat trout (Oncorhynchus clarkii) populations above knickpoints. Model results and the isolated O. clarkii populations provide implications for the transmission of erosion through basins and the resiliency of small isolated fish populations.
1.2. Introduction

Climate and tectonics are primary drivers of landscape change, yet the rates and mechanics of transient erosion signals initiated by these perturbations remain unresolved [Larsen and Montgomery, 2012; Whitaker and Boulton, 2012], and are likely to vary by geographic and tectonic setting. These signals are expressed as waves of accelerated channel and hillslope erosion that propagate through basins and can reside in a landscape after the climate and tectonic conditions have changed [Whipple and Tucker, 1999; Crosby and Whipple, 2006; Willenbring et al., 2013]. At the scales of drainage basins and mountain ranges, aspects of climate and tectonics can be deciphered in the topography and topology of bedrock rivers especially where streams incise at the front of erosional waves [Stock and Montgomery, 1999; Gasparini and Whipple, 2014; Willet et al., 2014]. Other factors, including the erodibility of bedrock, modulate the sensitivity and lead to the complexity of interpreting fluvial records of both climate and tectonics.

Stream incision, $\delta z/\delta t$, as it relates to climate and tectonics can be modeled with the following relationship:

$$\frac{\delta z}{\delta t} = U(x, t) - K(x, t)Q(x, t)^m \left| \frac{\partial z}{\partial x} \right|^n$$

(1)

where $z$ is elevation, $t$ is time, $U$ is rock uplift rate, $x$ is stream-long distance, $K$ is the erodibility coefficient, $Q$ is stream discharge, and $m$ and $n$ are non-universal positive exponents. In regions where stream incision, climate, and tectonics are relatively understood, it stands to reason that signals of transient erosion can be evaluated by parameterizing the above stream power model with the boundary conditions that are hypothesized to drive transience in a region. Many investigations in certain
settings assume topographic steady state with uniform $U$ and $K$, which yields this simplification of equation (1):

$$\frac{\Delta z}{\Delta x} = S = \theta A(x)^{-\theta}$$

(2)

where $S$ is channel slope, $A$ is drainage area as a surrogate for stream discharge, $\theta$ is the concavity index that is equal to $m/n$, and $k_s$ is the steepness index that is equal to

$$k_{sn} = \left(\frac{U}{k_{sn}}\right)^{1/n}$$

(3)

where $\bar{P}$ is the mean upstream precipitation [e.g., Gasparini and Whipple, 2014]. Equation (2) indicates that areas that demonstrate a power law relationship within a range of $\theta$-values are in steady state [Pederson and Tressler, 2012]. Channels that do not demonstrate a power law relationship retain transient signals embedded within their profile, given that the absence of a power law relationship cannot be attributed to significant variations of bedrock erodibility or precipitation.

Oversteepened stream reaches (knickpoints) can be the transient portion of streams and express climatic and tectonic signals [Neimann et al., 2001; Crosby et al., 2007]. Knickpoints result from many mechanisms, including bedrock erodibility contrasts, fault throw, or eustatic base level fall [Stock and Montgomery, 1999; Whitaker and Boulton, 2012]. The focus of this investigation are knickpoints that originate as fluvial hanging tributaries, which form in response to gradual or effectively instantaneous incision on the trunk river that outpaces incision along its tributaries [Wobus et al., 2006a; Crosby et al., 2007]. Intensified stream incision can be initiated by changes in relative base level, sediment load, and uplift—all of which are modulated by climate
Identifying knickpoints and transient erosion signals is complicated by noise in topographic datasets. This can be largely circumvented using an integral approach to stream longitudinal profile analysis developed by Royden et al. [2000] and Perron and Royden [2013]. This approach calculates the variable $\chi$ (chi), which incorporates upstream drainage area in an integral transformation of the $x$-coordinate of the longitudinal profile of a stream. Variables in equation (2) are separated and integrated along the $x$-coordinate [e.g., Royden et al., 2000; Perron and Royden, 2013]:

$$\int dz = \int \left( \frac{u(x)}{K(x)A(x)^m} \right)^{1/n} dx$$  

(4)

The profile is integrated from the chosen base level of the profile being analyzed, $x_b$, to points along the stream, $x$:

$$z(x) = z(x_b) + \left( \frac{u(x)}{K(x)A(x)^m} \right)^{1/n} \chi$$  

(5)

The integrand is ensured to be dimensionless by introducing a reference drainage area, $A_0$:

$$\chi = \int_{x_b}^{x} \left( \frac{A_0}{A(x)} \right)^{m/n} dx$$  

(6)

The reference drainage area is introduced to simplify calculations and the selection of the value does not effect interpretation. Plots of $\chi$ versus channel elevation [“chi plots” in Perron and Royden, 2013] can be used to identify knickpoints or knickzones as upstream migrating mathematical identities [“slope patches” in Royden and Perron, 2013].
This study aims to identify the contributions of climate and tectonics in injecting transient stream incision signals in the Clearwater River basin in the Olympic Mountains of Washington State. Pazzaglia and Brandon [2001] provided compelling evidence that this region is in flux steady-state over time periods that integrate glacial and interglacial cycles [e.g., Willett and Brandon, 2002]. During glacial intervals, the Clearwater River favored horizontal versus vertical bedrock incision of its streambed, forming a planar bedrock surface called a strath [Pazzaglia and Brandon, 2001]. Rock exhumation along the Cascadia subduction wedge drove these straths upward. As hillslope sediment supply to the channel network decreased at the end of glacial periods and discharge increased at the onset of local warming phases, the Clearwater River incised into these strath surfaces, stranding them as strath terraces from which stream incision can be reconstructed through time. The rates of exhumation from apatite fission track thermochronology and stream incision are similar at the 10^5-yr scales. Climate change drives fluctuations in the longitudinal profile of the river, although it is expected that the concavity of the Clearwater River’s longitudinal profile returns to the same profile in cycles >10,000 to 100,000 years [Pazzaglia and Brandon, 2001].

We hypothesize that river incision at the onset of a warmer period approximately 140 ka (Oxygen Isotope Stage (OIS) 6–5 transition) initiated a wave of accelerated erosion that continues to translate up the Clearwater River basin. During the time scale of this investigation, this river and its tributaries are entirely fluvial and there is no evidence for glacial activity in the basin. Our objective in this contribution was to determine if these hanging tributaries are instead a transient signal initiated by climatically or tectonically-driven adjustments to Clearwater River longitudinal
profile. This will allow us to consider how landscapes transmit erosion signals, the erosional processes (e.g., stream incision, mass wasting) that drive these signals, and the impacts of climate and tectonics on landscape domains in settings with transient features.

### 1.3. Field Setting

The Clearwater River is a tributary of the Queets River that is situated on the western slope of the Olympic Mountains above the Cascadia subduction wedge (Figure 1). This range emerged above sea level approximately 18 Ma as the Juan de Fuca plate began to subduct underneath North America. The Olympic Mountains achieved flux steady-state by ~14 Ma meaning that material accreted to and eroded from the wedge are balanced over long timescales [Brandon and Vance, 1992; Brandon et al., 1998; Pazzaglia and Brandon, 2001]. Frontal accretion drives the wedge westward relative to a fixed Puget Sound [Pazzaglia and Brandon, 2001; Stewart and Brandon, 2004].

The evidence of steady state was largely provided by a flight of stream terraces along the Clearwater River [Pazzaglia and Brandon, 2001]. Fluvial terraces were likely destroyed in adjacent valleys of the Olympic Mountains that experienced repeated late Quaternary alpine glacial advances, including in the adjacent Queets and Hoh River valleys (Figure 1). Six Pleistocene and Holocene terraces have been identified in the Clearwater River basin [Pazzaglia and Brandon, 2001; Wegmann and Pazzaglia, 2002]. Rates of incision and exhumation are similar along the trunk Clearwater River
as evidenced by the Pleistocene strath terraces, although the timescale over which exhumation balances incision along tributaries has not been addressed.

We hypothesize that deep stream incision into the strath of Quaternary Terrace 2 (Qt2) is contemporaneous with the onset of the erosional wave that propagated through the Clearwater River basin. The Qt2 strath was carved 180 to 140 ka, and buried 160 to 120 ka after which incision begun [Pazzaglia and Brandon, 2001]. Incision into the Qt2 strath and thickness of the supra-strath alluvial deposits of Qt2 are notably larger than the younger Clearwater terraces due to the intensity of the marine isotope stage (MIS) 6 glaciation [Chappell and Shackelton, 1986; Pazzaglia and Brandon, 2001]. Sediment supply was especially enhanced to the mid and lower Clearwater River by glacial outwash that spilled over the northern divide from the glaciated Hoh River valley [Thackray, 2001].

Basin-wide patterns of terrace elevations and glacio-fluvial deposits indicate that exhumation increases from the coast up the Clearwater River basin [Pazzaglia and Brandon, 2001]. The long-term erosion rate on the Pacific coast is about 0.3 m/ka and increases to 0.8 m/ka in the center of the Olympic Mountains [Figure 1c; Brandon et al., 1998]. The short-term erosion rates (~10–20 kyr) show a smaller inland increase from about 1.2 to 1.6 m/ka (Savage et al., 1991; Dragert et al., 1994). Precipitation follows a similar trend with the highest rates on the west side of the range due to orographic effects [PRISM, 2014]. Rates of bedrock incision and erosion are balanced when integrated over glacial to interglacial time scales (> 10 ka) [Pazzaglia and Brandon, 2001].
Figure 1. (a) Simplified tectonic setting of the Cascadian subduction wedge [modified from Pazzaglia et al., 2002]. Arrow indicates direction of plate convergence. HRF is the Hurricane Ridge Fault that separates accretionary wedge rocks of the Olympic Subduction Complex from structurally higher rocks of the North American plate. (b) Local relief map of the Olympic Mountains produced from a Shuttle Radar Topography Mission (SRTM) 30 m digital elevation model. Major rivers discussed in the text are labeled. (c) A grid of 30-year normal precipitation for the period of 1981–2010 is from PRISM [2014]. Exhumation contour map is from Brandon et al. [1998].
1.4. Methods

To identify patterns in stream incision, exhumation, and geomorphology, we modeled knickpoint propagation to test our hypothesis that river lowering at the onset of an interglacial period initiated a wave of accelerated erosion that remains expressed in the modern day Clearwater River basin. A conceptual model of this hypothesis is presented in Figure 2. A United States Geological Survey digital elevation model (DEM) provided inputs to identify knickpoints and model their propagation in tributaries of the Clearwater River basin. The National Elevation Dataset DEM has a horizontal resolution of 10 m and the root mean square error is 1.55 m [Gesch et al., 2014]. The age of knickpoints is evaluated using genetics of isolated fish populations collected from above tributary waterfalls in this basin.

1.4.1. Tributary selection and knickpoint identification

We limited our analyses to 14 tributaries of the Clearwater River that met the following criteria: (1) a drainage area greater than 5 km$^2$ to eliminate small channels along the hillslopes of the main river, (2) a trunk-to-tributary drainage area greater than 10:1 because Wobus et al. [2006a] recognized that the contrast of stream power at junctions with this drainage area ratio is necessary to form hanging tributaries, and (3) above valley km-23 because Pazzaglia and Brandon [2001] identified a structural perturbation that warped the Qt2 terrace below this location, which injects complications into our modeling effort (Figure 3). These tributaries were analyzed from their mouths to a drainage area threshold of 0.5 km$^2$. 
**Figure 2.** Conceptual model of a transient topographic signal originating from a decrease in the elevation of the trunk stream. In Time 1, graded streams have straight lines in chi-elevation space. As the trunk stream lowers in Time 2, hanging tributaries form. This creates two zones in chi-elevation space that are coincident with and separated by the locations of tributary knickpoints. In Time 3, upstream knickpoint propagation has occurred. This results in an increased rate of hillslope landsliding that drives a steepened hillslope zone towards the ridgelines. Inflections in chi-elevation space are now located at a higher χ-value.
The decision to use chi plots to identify the transient portion of the Clearwater River basin was motivated by noise in the DEM and potential variability of $K$ caused by differences in lithology and precipitation. $\chi$ was determined using equation (5) along the 14 selected tributaries with values extracted from the DEM with a program described in Mudd et al. [2014]. We used the following parameter values that are described in Mudd et al. [2014]: $A_0 = 1$ km$^2$, mean skip value = 3, minimum segment length = 20, number of target nodes = 100, and $\sigma = 3$ m, which is approximately 2 times the DEM elevation error. Channels were analyzed at drainage areas above 1 km$^2$ because this is the approximate location of the slope break in drainage area plots, indicating that channels upstream from this threshold are dominated by debris flows, and thus should not be included in an analysis centered on stream incision. Knickpoints were identified using chi plots and were placed at inflection points in chi plots of tributaries similar to the inflection point illustrated in the chi plots in Figure 2.

The most likely $m$ and $n$ ratio, or $\theta$, was identified by determining a cumulative goodness-of-fit criteria for the Clearwater River [Mudd et al., 2014]. This criteria, Akaine Information Criteria (AIC), determines the goodness-of-fit of a model while considering the complexity of the model where the minimum AIC indicates the best model [Akaine, 1974]. We tested a range of $\theta$-values from 0.2 to 0.95 every 0.05. The $\theta$-value with the minimum AIC for the Clearwater River was $0.55 \pm 0.03$ and was used in the model used in this study.
Figure 3. The tributaries of the Clearwater River included in this study are designated 1–14. Qt2 incision was measured along the same valley profile line of Pazzaglia and Brandon [2001]. Collection locations of cutthroat trout tissue is denoted by the black outlined reaches on tributaries 2, 3, 5, and 6.

1.4.2. Stream Power Model

An analytical model was used to explore the existence of a resolvable record of transient incision contained in tributary longitudinal profiles. In this modeling approach, knickpoints were formed where tributaries enter the main Clearwater River valley. This follows the hypothesis that knickpoints were generated as hanging tributaries that formed at the time when the Clearwater River was incising
into the Qt2 strath. This model was used to evaluate the distance that tributary knickpoints could travel over the time since Qt2 incision began. The StreamProfiler program was also used in this component of the study to automate data extraction from the DEM, extract channel data at 5 m elevation contours, smooth the profiles over a 100 m window, and calculate $k_{sn}$ along tributaries [Wobus et al., 2006b].

Knickpoint velocity across each cell was modeled along the selected tributaries following the approach of Rosenbloom and Anderson [1994]:

$$\frac{\delta x}{\delta t} = KQ^m S^{n-1}$$

The area exponent, $m$, was determined by identifying the mean $\theta$ in equation (2) for the fluvial portion of tributaries above the knickpoints identified with the chi plots, and solving for $m$ given the assumed value of $n$. The channel slope exponent, $n$, was assumed to be one, which is a common assumption for detachment-limited channels. When $n$ is equal to 1, equation (6) takes on the form:

$$\frac{\delta x}{\delta t} = KQ^m$$

Equation (8) is equivalent to the knickpoint celerity model used by others where the coefficient and exponent that yields the best fit of predicted and observed knickpoint locations were used [Crosby and Whipple, 2006; Berlin and Anderson, 2007; Gallen et al., 2013]. An important component in this study is that by having constraints upon stream incision from prior studies allows us to empirically constrain the erodibility coefficient, $K$, so that we do not have to appeal only to best-fit model results.
Given that the intent of the model is to determine incision below the transient knickpoints and that the modern day Clearwater River is at, or is approaching the steady-state profile, $K$ can be calculated by rearranging equation (3):

$$K = U k_{sn}^n \bar{P}^{-m},$$

(9)

where $k_{sn}$ was determined for each tributary below knickpoints. We used exhumation and precipitation grids for $U$ and $\bar{P}$, respectively. In the steady-state Olympic Mountains, $U$ was quantified as the exhumation rate that has been previously measured using fission-track thermochronology by Brandon et al. [1998]. The contour map of exhumation rates produced by Brandon et al. [1998] includes the Clearwater River basin (Figure 1c). Precipitation was extracted from a grid of 30-year precipitation normals (1981–2010) from PRISM [2014] that has a horizontal resolution of 30 arcseconds.

The discharge values, $Q$, required to predict knickpoint velocity with equation (8) were approximated from the PRISM [2014] precipitation grid. Precipitation was routed throughout the 10 m DEM. Stream discharge, $Q_j$, at each cell, $j$, is modeled with the relationship:

$$Q_j = \sum_{i=1}^{j} P_i a_i$$

(10)

where $P_i$ is the rainfall rate at location $i$ in the channel network, $a_i$ is the cell area at location $i$ in the channel, and $i$ includes all upstream cells. This assumes that all precipitation is converted to runoff as a simplification similar to other recent modeling efforts [Gasparini and Whipple, 2014; Han et al., 2014].
1.4.3. Model Experiments

Modeled knickpoints migrate to upstream cells following equation (8) until the hypothesized time period of 140 ± 20 ka is reached. Two sets of experiments were conducted (Table 1). The intent of the first set was to model the upstream distance knickpoints traveled given mean values of inputs (X1) and the uncertainty of the timing of Qt2 incision, the rates of exhumation and precipitation, and the value of $K$ (X2 and X3). In X2 and X3, the minimum or maximum of input values in each tributary basin were selected such that the range of travel distances were modeled. The modeled upstream travel distances were evaluated by comparing the location of the predicted knickpoints with the knickpoints identified with the chi plots.

In the second set of experiments, the effect of uplift and precipitation patterns on the final modeled knickpoint locations was explored. The duration was 160 ka and the maximum value of $K$ in each tributary basin was used for all experiments of this set. The patterns of uplift and precipitation values varied by experiment: combinations of spatiotemporally constant and gradient values (X4 and X5) and constant values for both uplift and precipitation (X6). The values of the gradient pattern were the mean uplift and precipitation for each tributary basin determined with data from Brandon et al. [1998] and PRISM [2014], respectively. The uniform values were the mean uplift and precipitation in the entire Clearwater River basin.
Table 1. Experiment parameter

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Model duration (ka)</th>
<th>Precipitation(\text{a}) pattern(\text{f}) value(\text{g}) (m/yr)</th>
<th>Rock uplift(\text{c}) pattern(\text{f}) value(\text{g}) (mm/yr)</th>
<th>(K^d) pattern(\text{f}) value(\text{g})</th>
<th>(m^e)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Knickpoint travel</td>
<td>X1 140</td>
<td>G mean</td>
<td>G mean</td>
<td>V mean</td>
<td>0.55</td>
</tr>
<tr>
<td></td>
<td>X2 120</td>
<td>G min</td>
<td>G min</td>
<td>V min</td>
<td>0.53</td>
</tr>
<tr>
<td></td>
<td>X3 160</td>
<td>G max</td>
<td>G max</td>
<td>V max</td>
<td>0.57</td>
</tr>
<tr>
<td>Uplift and precip</td>
<td>X4 160</td>
<td>G mean</td>
<td>U 0.42</td>
<td>V max</td>
<td>0.57</td>
</tr>
<tr>
<td>patterns</td>
<td>X5 160</td>
<td>U 3.5</td>
<td>G mean</td>
<td>V max</td>
<td>0.57</td>
</tr>
<tr>
<td></td>
<td>X6 160</td>
<td>U 3.5</td>
<td>U 0.42</td>
<td>V max</td>
<td>0.57</td>
</tr>
</tbody>
</table>

\(^a\)Age of Qt2 from Pazzaglia and Brandon [2001].
\(^b\)From PRISM [2014].
\(^c\)From Brandon et al. [1998].
\(^d\)Determined using equation (9).
\(^e\)Fluvial portion of tributaries in the relationship, \(S = k_s A(x)^{-m/n}\).
\(^f\)G=gradient, U=uniform, V=variable.
\(^g\)Statistic values (mean, min, max) were calculated for the basin of each tributary.
\(^i\)Precipitation and uplift inputs varied by watershed for the knickpoint travel distance experiments.
\(^k\)Precipitation and uplift inputs were the same throughout the Clearwater River basin in the uplift and precipitation patterns experiments.

1.4.4. Geochronology

Fish genetics provides a chronometer independent of our knickpoint modeling and terrace dating methods used in Pazzaglia and Brandon [2001]. Within the premise of our hypothesis, aquatic organisms would become isolated above knickpoints once waterfalls reach an impassable height. Waterfalls with a height of 2 m have been sufficient to genetically isolate cutthroat trout populations [Gresswell et al., 2006]. Knickpoints then block upstream passage of fish, which instigates genetic drift and decreases population genetic variation [Wofford et al., 2005]. Introduction of alleles—
alternative forms of a gene—to fish populations upstream of knickpoints is then limited to mutations, which along with the genetic mutation rate of a species, operates as a “molecular clock” that records the time since knickpoint formation. Results of genetic analyses can then be used as an independent chronometer to gauge the fidelity of predicted knickpoint initiation in the model experiments.

We collected and analyzed DNA from coastal cutthroat trout (*Oncorhynchus clarkii clarkii*) specimens above knickpoints with heights that exceeded 2 m to assess the genetic distance of subpopulations and to estimate the time since these populations were connected. This species of fish has existed in the Cascadia region south of the limit of Quaternary glacial ice sheet advances for millions of years [McPhail and Lindsey, 1986]. Many headwater populations of *O. clarkii clarkii* are commonly the only fish species found above waterfalls in the Coast Ranges of Oregon, the Cascades, Olympic Mountains, and Clearwater River [e.g., Edie, 1975; Guy et al., 2008].

DNA material was extracted from Cutthroat trout on tributaries 2, 3, 5, and 6 (Figure 3). The collection sites on Tributaries 2, 3, and 5 were above waterfalls >10 m high that completely block upstream passage of fish. The collection site on tributary 6 was below a waterfall, but in a broad knickzone; therefore fish from tributary 6 have the potential to be reproductively connected with populations in the main river. DNA samples were collected from 15 to 25 individuals at each site. A total of 72 individuals were sampled.
The adipose fins of Cutthroat trout were separated from individuals in the field, and the fin was stored in a 90% ethanol solution until it was analyzed in the lab. DNA was isolated from adipose fin tissue using the Nucleospin Tissue Kit [Macherey-Nagel Inc., Bethlehem, PA, USA]. DNA concentrations were standardized to no more than 50 ng/µL using a Nanodrop 1000 spectrophotometer. All DNA samples were genetically verified as Coastal Cutthroat Trout by generating DNA barcodes and querying against the online BOLD database [http://www.boldsystems.org]. DNA barcodes (derived from the mitochondrial COI gene) and DNA sequences from the ND2 gene (also from the mitochondrial genome) were prepared for the 72 individuals of *O. clarkii clarkii* using standardized protocols implemented by BOLD Systems [Ratnasingham and Hebert, 2007] for DNA barcodes and following protocols from Kocher et. al 1995 for the ND2 gene. Polymerase chain reactions (PCRs) were assembled as 10 µL reactions containing 5 µL GoTaq MasterMix [Promega, Inc., Madison, WI, USA], 0.01 mM final concentration of each primer cocktail following the protocols outlined in Ivanova et al. [2007] and using approximately 10-50 ng of genomic template. Successful PCR reactions were cleaned for sequencing using ExoSAP-IT [USB, Cleveland, OH, USA] following the manufacturer’s recommendations. DNA sequencing was performed on an ABI 3130XL Genetic Analyzer [Applied Biosystems, Foster City, CA, USA] following the manufacturer’s recommended protocols for DNA sequencing and cleanup. Sequences that were 1048 base pairs long were proofed and aligned using Biomatter’s Geneious v 6 software [http://www.geneious.com].
The widely used salmonid mtDNA divergence rate of 1% My\(^{-1}\) was used to estimate the time since populations became isolated from the main river [Smith, 1992], which we assume to have been synchronous with knickpoint formation.

1.5. Results

1.5.1. Knickpoints identified using the DEM

Two broad trends are observable in the chi plot gradients versus elevation for the 14 tributaries (Figure 4). On all tributaries, the downstream trend is steeper than the upstream. Elevations at which these trends converge are coincident with either tributary waterfalls or broad knickzones, and are within 500 m of the maximum \(k_{sn}\) on each tributary. The two uppermost tributaries demonstrate only one broad trend indicating that the entirety of these streams can be characterized as steady-state profiles [Perron and Royden, 2013]. The point where the two trends meet in the other tributaries will henceforth be referred to as knickpoints for simplicity even though some of these reaches could be called knickzones. Knickpoints were not identified on the two uppermost tributaries that resemble steady-state profiles; therefore knickpoints were not modeled on these streams. The elevation of knickpoints on the tributaries that do not exhibit steady-state profiles ranges from 98 to 432 m above mean sea level. These knickpoints are less than 100 m above the Qt2 strath terrace that outcrops in the main valley nearest to the tributary outlet (Figure 5).

Vertical velocities of knickpoints were calculated as the elevation difference between the predicted knickpoint locations and the elevation of the Qt2 strath nearest to the tributary outlet over 120, 140, and 160 ka. The resulting velocities are 0.07–1.0 mm/yr.
and they increase up the basin at rates similar to that observed from the rock exhumation and terrace incision data from the Clearwater River (Figure 6a).

**Figure 4.** Chi plot of representative tributaries and the Clearwater River (CR). Not all tributaries are shown for clarity. Tributary knickpoints are identified by elevation where there are breaks in linearity in the chi-elevation plot. The entirety of tributary 13 and 14 (not shown) were linear in chi-elevations space, so knickpoint locations were not defined for these streams.
Figure 5. Stream longitudinal profiles of the Clearwater River (CR) and study tributaries (1–14). Profiles were smoothed as described in the text. Point locations of the Qt2 strath terrace along the valley profile of Pazzaglia and Brandon [2001] were projected onto the Clearwater River profile. Strath elevation measurement error is within the point symbols. The profile for Tributary 3 is shown as a gray line only to differentiate it from the other tributary profiles on the plot.
Figure 6. Qt2 incision rates and knickpoint vertical velocity. (a) The rate of Qt2 incision determined by Pazzaglia and Brandon [2001] assumes incision occurred along the valley profile instantaneously at 140 ka. Gray error bars report the uncertainty associated with the measurements of strath elevation. The black error bars report the range of potential vertical velocities (given the travel time of 140 ± 20 ka) of tributary knickpoints identified with chi plots. (b) Qt2 incision is shown for comparison with the vertical velocities of predicted knickpoints. X2 and X3 are not shown because these experiments follow the same trend as X1.
(a) Incision of Qt2 strath (140 ky)

Observed knickpoint vertical velocity

(b) Predicted knickpoint vertical velocity

- X3 (uplift and precipitation gradient)
- X4 (uniform uplift, precipitation gradient)
- X5 (uplift gradient, uniform precipitation)
- X6 (uniform uplift and precipitation)
1.5.2. Model Experiments

Modeled stream discharge and rock erodibility are within expected ranges. Modeled stream discharge, $Q$, at the outlet of the Clearwater River is within the range of discharge measured by a stream gauge during the same period (1981–2010) of the PRISM 30-year precipitation normals. The stream gauge data is normalized by upstream drainage area to compare the discharge values. Normalized discharge on the Clearwater River is $1.41 \pm 0.289 \times 10^9$ m$^3$/yr and modeled discharge was $1.23 \times 10^9$ m$^3$/yr. The erodibility coefficient, $K$, is between $1.1 \times 10^{-3}$ and $5.0 \times 10^{-4}$ m$^{0.16}$ yr$^{-1}$, which is within the range of values reported elsewhere [Snyder et al., 2000].

Overall, predicted knickpoint travel distance was similar to the observed distance between knickpoints and tributary outlets (Figure 7). Knickpoint travel distance was under-predicted in the X2 experiment when exhumation, precipitation, and the timing of Qt2 incision were averaged. However, the maximum of these factors approaches a 1:1 fit of observed and predicted knickpoints in X3. Predicted knickpoint locations were within the bounds of the three trials with two exceptions (Figure 8). The knickpoints of tributaries 3 and 4 were predicted outside of the predicted bounds by less than 50 m upstream of the terminal position in X3. This could be attributed to the precipitation and/or exhumation gradient not being fully represented by the inputs grids, or the cumulative effect of minor variations in rock hardness and channel geometry. These residuals, knickpoint travel distance, and their velocity do not follow clear trends with respect to the observed knickpoint locations. As predicted by theory, knickpoints do reach higher elevations further up the basin when comparing drainage basins of similar size (Figures 6, 8).
The vertical velocity of predicted knickpoints is within the range of Qt2 incision (Figure 6b). Velocity was lowest in experiments with uniform uplift (X4 and X6) and highest in experiments with an uplift gradient (X3 and X5). Uplift is incorporated in the model through the calculation of the erodibility coefficient, $K$ in equation (9). Precipitation is incorporated in $K$ as well as $Q$ in equation (10). The value of precipitation is larger than uplift, although it is raised to $m$, an exponent less than 1, in both calculations. The range of velocity in experiment X6 (uniform precipitation and uplift) was smallest (<0.2 mm/y). The effect of averaging precipitation and uplift is increasing vertical velocity downstream and decreasing it upstream from the mean input values in X1.
Figure 7. Comparison between observed distance between knickpoints and tributary outlets and modeled knickpoint travel distance. Linear regressions show best-fit line for the experiments.
Figure 8. Map view of modeled knickpoint locations. Inset shows the symbology used for each tributary, including the terminating locations of knickpoints in experiments, X1–X3. The distance and horizontal velocity that a knickpoint traveled over X2 is shown by the bolded channel line where the color indicates the knickpoint horizontal velocity at each cell.

1.5.3. Fish Genetics

Mitochondrial DNA of *O. clarkia clarkii* populations above knickpoints differ by one to three base pairs amongst tributaries (Figure 9). When these populations are compared to the main river, 9 to 10 base pairs differ. Given that the sequences are composed of 1048 base pairs, populations diverge by 0.095–0.286% amongst tributary populations. This equates to the timing of divergence of 95–286 ka, assuming a 1% my⁻¹ divergence rate [Smith, 1992].
Figure 9. Comparison matrix of mtDNA base pair differences between *O. clarkii clarkii* populations. Each tributary population is compared to each other population in this matrix. Histograms show counts of base pair differences in fish-to-fish comparisons. The locations of the fish individuals sampled are shown on Figure 3.

1.6. Discussion

The majority of the Clearwater River is at or approaching its steady-state profile, although the signal of at least one perturbation is retained in tributaries and leads the front of an erosional wave that is propagating up the basin. This is indicated by the similar position of inflection points (knickpoints or knickzones) in chi-elevation space, and the fidelity of experiments X1–X3 relative to the observed knickpoints.
The positions and elevations of knickpoints relative to the Qt2 terrace follows our hypothesis that knickpoints formed as fluvial hanging valleys above the main Clearwater River. In certain conditions, knickpoints on hanging tributaries migrate upstream causing hillslopes to adjust to a lowered local base level. This phenomenon has been described by some authors [e.g., Bigi et al., 2006; Crosby et al., 2006; Gallen et al.; 2011] as an erosional wave characterized by an increase in the stream incision rate and landsliding frequency. In the Clearwater River basin, an increased frequency of landsliding below knickpoints along tributaries is supported by an interpretation of sediment provenance by Belmont et al. [2007]. With an analysis of $^{10}$Be concentration in alluvial sand and gravel throughout this basin, these authors proposed that the portion of tributary 6 below the knickpoint is subject to an increased frequency of landsliding compared to the upper portion and that this relationship is evident in comparisons of millennial-scale $^{10}$Be basin average erosion rates from above and below the knickpoint. If this spatial relationship of landsliding in this tributary is applicable to the other tributaries on the Clearwater River, this supports our notion that hillslopes are adjusting to the erosional wave initiated by Qt2 incision.

In the conceptual model presented in Figure 2, fish would be isolated above knickpoints. The mutation of mitochondrial DNA of Cutthroat trout populations in the Clearwater River basin is coincident with the onset of Qt2 incision. The time span that these populations were separated is wide (95–286 ka) and the assumption that fish populations began with no genetic variation cannot be evaluated. Genetic variation is minor in this basin, which is typical for this species, but the DNA sequences of tributary populations above knickpoints do cluster, indicating that the
individual fish sampled were geographically established for a period of time sufficient to generate characteristic sequences. Nevertheless, our interpretation of the timing of knickpoint formation is not discounted by the genetics of the sampled fish populations given that the timing of population divergence (95–286 ka) does envelope the timing of Qt2 strath incision (140 ± 20 ka). Genetic divergence was quantified as the number of nucleotide differences between populations. Although, the technique described in this chapter does not adequately consider the complexity of genetic mutation. Craw et al. [2008] used the program, IMa to determine the timing of stream basin capture with the genetic mutation of galaxiid fish.

Knickpoints are implicit consequences of fluvial hanging tributaries [Crosby et al., 2007], their temporal persistence are reflected in the $n$-value (exponent of slope) of the channel, and they are theoretically degraded when $n$ is not equal to one [Royden and Perron, 2013]. We assumed a value of one for $n$ throughout the Clearwater River basin. Mudd et al. [2014] used a more targeted approach to identify the likely range of $n$ by analyzing $\chi$ relative to uplift. By further exploring this method, $n$ can be determined where erosion constraints are known in the Clearwater River basin. Using the approach of Mudd et al. [2014] here could improve the parameterization of the stream power model and facilitate a knickpoint persistence analysis.

Theory and observation predicts that in detachment limited fluvial systems with little spatial variability in uplift and erodibility pulses of stream incision will translate through basins at a similar vertical rate that is approximately equal to the new uplift rate. [Niemann et al., 2001; Schildgen et al., 2007], meaning that knickpoints in these situations exist at approximately the same elevation (isoelevation
knickpoints). However, knickpoints in the Clearwater River basin are observed and predicted in a wide range of elevations. Uplift was assumed to be uniform in regions where isoelevation knickpoints have been identified, [e.g., Niemann et al., 2001; Gallen et al., 2011]. Berlin and Anderson [2007] attributed the range of knickpoint elevations in the Roan Plateau to the complex uplift history of this region, while Gallen et al. [2013] ascribed the isoelevation of tributary knickpoints in the Cullasaja River basin of the southern Appalachians to pulses of basin-wide incision driven by multiple base level lowering events affecting the main channel.

In the Clearwater River basin, the non-isoelevation knickpoints that initiated as fluvial hanging valleys are maintained by gradients of exhumation and precipitation. These similar rates demonstrate how erosional signals can propagate in basins with exhumation and precipitation gradients. The vertical velocity that would position knickpoints in their present locations follows the long-term (>10^5 yr) magnitude of exhumation and stream incision (Figure 6).

An exceptional opportunity to investigate the mechanisms of topographic transience is provided by previous work in the Clearwater River that quantified the history of stream incision and exhumation. The Olympic Mountains are asymmetric from west to east. Subduction of the Juan de Fuca plate leads to a sharp exhumation gradient on the eastern slope of the Olympics Mountains where hillsides and channels are steep (Figure 1). This asymmetry suggests a disequilibrium across the major drainage divide that trends north-south [Willett et al., 2014], which in the case of this region, would drive this range eastward. Using numerical models, Willett [1999] proposed that orographic precipitation also plays a role in this asymmetry by
enhancing erosion on the west side of this range. Both increased exhumation and precipitation enhance fluvial incision, which in the case of the Clearwater River basin, compels topographic transience at a faster rate where these drivers reach the most rapid rates. The Clearwater River is exceptional in this region because it has not been recently glaciated like the larger Hoh and Queets basins to the north and south, respectively, where montane glaciers destroyed terraces, knickpoints, and potentially other indictors of transient erosion signals initiated as fluvial hanging valleys. These signals may be attenuated or non-existent in valleys routinely scoured by montane glaciers.

The presence of a transient wave of accelerated stream incision in the Clearwater River basin may explain the poor predictive ability of incision models conducted by Tomkin et al. [2003] on the trunk stream of this basin. Our efforts were focused upon the tributaries of this river, although broad knickzones along the trunk river share similarities with the tributaries in chi- and distance-elevation space (Figures 4, 5), which follows some models of fluvial hanging valley formation [Crosby et al., 2007]. Tomkin et al. [2003] tested six bedrock incision models along the Clearwater River, and none of these models successfully predicted the downstream decrease in incision. Fluvial hanging valleys are not well explained with stream-power based river incision models [Crosby et al., 2007], and further complicated in regions where precipitation and exhumation significantly vary. We did not attempt to reformulate stream incision models to dynamically incorporate the formation of fluvial hanging valleys throughout the basin. We instead tested the potential of hanging valley formation by assuming this was the case, using existing models and constraints provided by known process rates and geomorphology, and comparing model
results with present-day topography. Future studies could combine the approaches of Tomkin et al. [2003] along the trunk river and the present study along tributaries to reformulate stream incision models to incorporate the effects of stream power contrasts at trunk-tributary junctures subsequent to a change in boundary conditions.

1.7. Conclusions
Stream incision into a Pleistocene strath terrace initiated a wave of accelerated erosion that begun to propagate up the Clearwater River basin 140 ± 20 ka. The physical location of tributary channel knickpoints demonstrate that climatic and tectonic perturbation signals do not progress uniformly up basins where exhumation and precipitation are not uniform in the upstream direction. Rather, signals progress more rapidly where exhumation and precipitation are greatest. These findings support other recent studies that propose that topographic adjustment to climate and tectonic perturbations can span millennial-scale climate fluctuations. Lastly, fish genetics as a geochronologic tool shows promise in investigating questions regarding landscape change, especially those related to stream incision and capture. When planning to use this tool, it is vital to select appropriate genetic markers (i.e. mitochondrial, microsatellite) that are able to resolve the timescale in question.
1.8. References


Chapter 2

The Hydrometeorological Conditions of Erosion Along a Streambank Comprised of Anthropogenic Sediment

2.1. Abstract

Streambank erosion is a primary source of suspended sediments in many waterways of the Atlantic Piedmont, eastern U.S. This problem is exacerbated where banks are comprised of fine sediment produced by the intensive land use practices of early European settlers. A stream in this region, Richland Creek incises laterally into banks comprised of three stratigraphic layers associated with historic land use: pre-European settlement, early European agriculture and development, and water-powered milldam operation. This stream is episodically turbid following precipitation, especially along incised reaches with exposed bank sediments.

The combined hydrometeorological conditions and dominant processes driving streambank erosion were investigated with analyses of stream stage, precipitation, and streambank face topography over 19 months between 2010 and 2012. Digital terrain models (DTMs) of the streambank were created from nine terrestrial lidar scanning surveys. Changes in the streambank face and volumetric flux were quantified by differencing the DTM time-series and analyzing a 3-dimensional voxel model, respectively. The spatial variability of erosion during a simulated precipitation event was examined in a field-based experiment.
In both the monitoring period and field-based experiment, the largest volume of erosion occurred where sediment columns detached along vertical desiccation and horizontal seepage cracks. The latter were most prevalent at the base of the sandy sedimentary layer deposited during early European agriculture and development. Slumping and collapse of sediment columns composed mostly of over-lying fine-grained mill pond sediment were more frequent in summer months when precipitation was most intense. This created the situation where the volume of streambank erosion was largest during summer months, contrary to the dominance of erosion in winter months observed in many other locations. This study demonstrates that anthropogenic sediment produced in historic times alters not only the volume of fine sediment that enters streams but also the process that dominates streambank erosion.

2.2. Introduction
Suspended sediment is the primary nonpoint-source pollutant in streams throughout the U.S. Atlantic Piedmont despite improved soil conservation practices (Neary et al., 1988; US Environmental Protection Agency, 2014). Significant lengths of stream reaches in this region are impaired for turbidity and nutrients (Deamer, 2009). Increasingly, river scientists are recognizing the role that historic land use practices have had in creating vast stores of sediment along streams, commonly referred to as legacy sediment (Trimble, 1974; Walter and Merrits, 2008; Stinchcomb et al., 2013; James, 2013; and references therein). Beginning in the 16th century, land clearing, agriculture, and development throughout the Atlantic Piedmont increased upland erosion rates 50–200 times above long-term geologic rates (Trimble, 1975),
and much of this legacy sediment remains stored in valley bottoms (e.g. Phillips 1992; 1993), is exposed along reaches of active streambank erosion, and can constitute a significant source of the suspended sediment load in streams (Jackson et al., 2005).

Naturally deposited sediment eroded from streambanks can contribute significant percentages of the total suspended sediment load of streams (Fox et al., 2007; and references therein). Streambanks composed of legacy sediment have especially high potential to contribute sediment to streams (Merritts et al., 2011; Wegmann et al., 2012). Despite the efficiency of streambank erosion and the increasing recognition of legacy sediment, little research has evaluated how hydrometeorological conditions effect the erosion of legacy sediments at the sub-reach scale.

Streambank erosion occurs by the detachment and removal of bank materials and is often categorized as fluvial, mass wasting, subaerial or groundwater seepage (Thorne, 1982; Lawler, 1994; Fox et., 2007). Fluvial erosion is the detachment and transport of bank material directly by stream flow. Mass wasting occurs (1) where the mid to upper bank is undercut by fluvial erosion or groundwater seepage (Chu-Agor et al., 2008; Fox and Felice, 2014), or (2) along failure planes created by subaerial processes (Green et al., 1999; Couper and Maddock, 2001). Subaerial processes, including soil desiccation and freeze-thaw, result from changes in soil moisture and hydrometeorological conditions that weaken the streambank soil (Wynn et al., 2008). Subaerial processes are often thought of as “preparatory” steps that increase bank susceptibility to erosion (e.g. Wolman 1959, Couper and Maddock, 2001); although Couper and Maddock (2001) argue that these processes
can directly cause sediment detachment and transport. Seepage erosion occurs where the groundwater table is ephemerally perched on a soil layer with low permeability, which causes rapid subsurface flow along a hydraulic gradient towards the stream that liquefies soil particles and induces mass wasting where the flow intercepts and undercuts the streambank (e.g. Fox et al., 2007).

Hydrometeorological conditions have an important temporal control on streambank erosion. Numerous studies have attributed rapid erosion rates in winter to freeze-thaw, which reduces granular interlocking in streambank soils when the zero-degree isotherm is frequently crossed (e.g. Wolman, 1959; Hooke, 1979; Wynn et al., 2008). Conversely, when temperatures remain above freezing, the drying of semi-saturated cohesive soils causes them to shrink, and desiccation cracks to form between soil blocks (Dietrich and Gallinatti, 1991; Green et al., 1999). Seasonal, prolonged wet periods rather than singular severe storms often lead to a greater reduction of interparticle forces that lead to more severe bank erosion (Wolman, 1959; Simon et al., 2000; Couper, 2003).

The primary objective of this research is to identify the hydrometeorological conditions and processes that dominate erosion of a streambank comprised of legacy sediment. Within the Atlantic Piedmont, we utilized a field location in Umstead State Park, which was established in 1934. Many streambanks within the park contain a well-documented sequence of anthropogenic sediment, including a reach of stream above a milldam that breached circa 1910 (Lewis 2011; Wegmann et al., 2012). Park land use restrictions during much of the last century has limited anthropogenic influence upon geomorphic processes, which makes this location
suitable to investigate modern day erosion of streambanks comprised of legacy sediment.

In this study, we assumed that the processes that have eroded the streambank can be identified by analyzing the morphology of the streambank. Terrestrial laser scanning (TLS) surveys were conducted over a 19-month monitoring period to quantify the volumetric and morphologic changes along a stream reach in Umstead State Park. Hypotheses on the dominant process that erodes anthropogenic sediment at this location were formed by comparing TLS data to stream stage and precipitation data. These hypotheses are then tested in a field-based experiment that was designed to induce streambank response to a simulated precipitation event. The findings of our study improve our understanding, and should inform modeling efforts aimed at tracing waterway pollution by suspended sediments and absorbed nutrients (e.g., N and P), both of which are critical to reducing the delivery of suspended sediment and absorbed nutrients to drinking water and recreational reservoirs and estuaries (e.g. Green et al., 1999; Paerl, 2009; Voli et al., 2013).

2.3. Field Setting

2.3.1. Legacy Sediment in the Atlantic Piedmont

Many streams across the Atlantic Piedmont incise into broad surfaces that are often described as the “valley flat” that were once assumed to be overbank silt and clay flood deposits of migrating, meandering streams (e.g. Wolman, 1955). In the twentieth and early twenty-first centuries, recent deep incision into the valley flats was predominately attributed to increased stream power due to urbanization
(Jacobson et al., 1986; Walsh et al., 2005). More recently, the formation of valley flats has been attributed to land use practices of early European settlers combined with stream modifications (e.g., the construction of tens-of-thousands of low-head dams) along Piedmont streams (Jackson et al., 2005; Walter and Merritts, 2008).

Many streams throughout the Atlantic Piedmont were not turbid at the onset of European settlement according to anecdotal sources (e.g. Lyell, 1849; Trimble, 1974 and references therein), indicating that minimal upland erosion occurred prior to early European settlement in this region. Depleted soil conditions and extensive hillslope erosion resulted from deforestation, one-crop production, farming on steep hillsides, and other intensive agriculture practices and land development (Costa, 1975; Jackson et al., 2005). This problem was exacerbated by Colonial to post-Reconstruction stream damming for water power. Tens of thousands of milldams were constructed across the Atlantic Piedmont by the beginning of the twentieth century (Walter and Merritts, 2008). Thick layers of sediment eroded from uplands were deposited in mill ponds, and while many of these dams have since disappeared, slack water deposits remain in valley bottoms throughout the Piedmont. Streams where milldams once operated now carry anomalously high amounts of suspended sediment that is entrained from these deposits (Walter and Merritts, 2008).

Numerous milldams existed in the upper Neuse River basin located in the Atlantic Piedmont (Figure 1a, 1b). An unpublished inventory of historic dam sites locates 83 in the upper Neuse River basin of North Carolina (Doug Swords, personal communication, 2012). This river is the third largest river basin in the State and
includes 5,470 km of freshwater streams, 230 km of saltwater streams, and 150,000 hectares of saltwater estuary. A significant percentage of the freshwater and estuarine stream kilometers in the basin are impaired due to elevated turbidity, reduced biological integrity, and low dissolved oxygen (Deamer, 2009).

2.3.2. Richland Creek and Cook’s Milldam

The study area is a reach of stream along Richland Creek, a 303(d) listed impaired tributary of Crabtree Creek and the Neuse River (Figure 1c). A milldam once operated on this creek 1.5 km upstream from the confluence of Richland and Crabtree Creeks (Wegmann et al., 2012). The drainage area upstream of the former milldam is approximately 25 km². The soils in this basin are deep (>2 m), moderately-to-well drained, and weathered from felsic gneiss and schist (Cawthorn, 1970).
Figure 1. (a) The study area is located in the Atlantic Piedmont physiographic province (AP) of North Carolina (NC). (b) Locations of historic milldams throughout NC. (c) The Richland Creek catchment upstream from Cook’s Milldam (gray outline). US Geological Survey stream gage locations (triangles) are shown for Crabtree (CSG) and Richland Creeks (RSG), and the State Climate Office of North Carolina rain gage (RG). (d) Section of Richland Creek where slack water deposits (beige area) created by Cook’s Milldam can be currently identified. Automated turbidity samplers deployed by Lewis (2011) are located up and downstream of mill pond deposits. Modern day flood control impoundment and reservoir is visible in the lower left corner of figure. (e) 2011 Aerial photo of the study reach, where Richland Creek flows between the surveyed streambank and the TLS scanner locations. The contour interval is 2 m.
According to municipal documents, Cook’s Milldam, which flooded approximately 0.1 km\(^2\) of valley bottom (Lewis 2011; Wegmann et al., 2012) was constructed in 1798 or shortly after and failed during a flood in 1910 (North Carolina Division of Archives and History, 1979). Hunt (2011) documented that trees growing on the floor of the former mill pond in the year 2010 had a mean age of 97 ± 3 years, suggesting that Richland Creek incised into the millpond sediments rather quickly following the 1910 breaching of the milldam. The foundation of the milldam is exposed at the elevation of the modern day stream. This indicates that (1) deep stream incision and the high streambanks in this catchment are due to Richland Creek returning to its pre-milldam elevation, and (2) the stream is no longer incising given that the stream has returned to its pre-milldam elevation. Instead the stream shifts laterally, which entrains sediment as the bank is eroded.

Mill pond slack water deposits are exposed immediately upstream of the milldam foundation where bank height rises from < 1 m to > 3 m above the stream. Wegmann et al. (2012) describes the characteristics and origins of three stratigraphic layers that comprise the banks of Richland Creek along the former mill pond reach (Figure 2). Base flow of Richland Creek intersects pre-European sediment (PES) that consists of quartz-rich pebbles and cobbles that interfinger with gleyed clays and silts interpreted as hydric wetland soil deposits. Unconformably above the PES is the early European legacy sediment (EELS) characterized by sand grading upwards into silty-clay. The EELS are thought to represent anthropogenically-driven increases in the rates of upland erosion and valley bottom deposition prior to the construction of milldams along a given stream following European settlement of an area. The upper streambank stratigraphic unit is mill pond legacy sediment (MPLS) that is composed
of fine-grained clayey silt interbedded with thin sandy flood event layers. MPLS is the sediment that was impounded in the slack water environment created by the construction of Cook’s Milldam.

Precipitation drives episodic increases in the turbidity of Richland Creek. Wegmann et al. (2012) documented the impact of precipitation and turbidity along this stream using automated turbidity samplers—deployed by Lewis (2011)—upstream and downstream of milldam deposits (Figure 1d). These samplers measured numerous turbidity increases in the downstream direction following precipitation, including a storm on February 4 and 5, 2011 (Figure 3). Turbidity at the downstream sampler was below background levels established by the upstream sampler until the 9th hour of sampling. Turbidity then exceeded 52 Nephelometric Turbidity Units (NTUs) by the 23rd hour of sampling, which is approximately five times higher than background levels, and is greater than the allowable standard of 50 NTUs established as the water column critical to protect aquatic life in all North Carolina fresh flowing streams (Deamer, 2009). Along the Cook’s Mill site on Richland Creek, steep increases of turbidity occur at approximately the same time of both rising limbs of the stream hydrograph (Figure 3). This is contrary to the often observed phenomena where streambank failure, which would increase turbidity, occurs on the recession limb of the hydrograph (Fox et al., 2007; Wilson et al., 2007).
Figure 2. Stratigraphic section of the scanned bank (location in Figure 1e; after Wegmann et al., 2012): (a) Photograph of a section of the bank that is representative of Richland Creek upstream from the former milldam. A crevice created by seepage erosion is visible at the contact of the pre-European and early European stratigraphic layers. The three layers are described in the text. (b) Generalized stratigraphy and grain-size variation of the three layers. The basal age of the pre-European layer was determined by Wegmann et al. (2012) with a single radiocarbon date. Other layers were constrained with historical records compiled by Hunt (2011).
The streambank bares the imprints of multiple erosion processes. Small cavities along the lower bank implicate aggregates entrained by fluvial erosion. The mid-to-upper bank is formed of elongated, sediment columns some of which have collapsed, which indicates that mass wasting is prevalent. Seepage cracks are also common, such as the crack visible at the base of the EELS in Figure 2a. These observations motivate our investigation into the processes that increase turbidity within this system.

Figure 3. Stream turbidity measured both up and downstream of the slackwater deposits related to the former Cook’s mill pond. Measurements were made with automated samplers during a storm event (after Lewis, 2001). Stream water samples were taken every hour for a 24-hour period. Turbidity measured during other precipitation events followed a similar trend. See Figure 1d for sampler locations.
2.4. Methods

The spatiotemporal patterns of erosion and process dominance were investigated at a reach of streambank 11.5 m long by 3.2 m high along Richland Creek (Figure 1e). This reach was selected because (1) prior research by Starek et al. (2013) was conducted at the same location, which enabled us to extend the time period we analyzed, (2) it is stratigraphically representative of streambanks along Richland Creek where legacy sediment is exposed, and (3) it was morphologically representative of erosional hotspots along Richland Creek. The section is on the cutbank of the stream, was minimally vegetated, and contained a large volume of sediment at its toe that had detached from the bank. By studying an erosional hotspot instead of a reach with erosion representative of the entirety of the stream, we can characterize the processes that detach a large portion of the total sediment that enters streams. Additionally, given the small scales we investigated, streambank change along reaches with minor amounts of erosion would not be detected outside measurement uncertainty.

Flux of the stratigraphic layers and hydrometeorological conditions were monitored over a 19-month period during 2010–2012. Next, the conditions that appeared to be driving sediment flux during the monitoring period were simulated in a field-based experiment. Over both periods, volumetric changes of stratigraphic layers were quantified using high resolution digital terrain models (DTMs) of the streambank (Starek et al., 2013). DTM were constructed from TLS surveys that were conducted periodically during streambank monitoring and the field-based experiment.
2.4.1. 19 months of Streambank Monitoring

The hydrometeorological conditions and erosional processes driving sediment to be detached from the streambank, transported along the bank, and entrained by stream flow were assessed with TLS surveys and precipitation and stream stage data of the Richland Creek area. Nine TLS surveys were conducted between December, 2010 and June, 2012. The 19-month monitoring period was divided into eight epochs where each epoch was bounded by two sequential TLS surveys. Monitoring epochs are henceforth referred to as Epochs 1 thru 8. The methodologies to acquire and process TLS data, and characterize flux using this data are described in later sections.

The metrics that were used to assess the hydrometeorological conditions of the study area were precipitation intensity, precipitation frequency, and stream stage. These metrics were compared to the volume and pattern of streambank erosion during monitoring epochs. Precipitation intensity and stream response were analyzed as the 98th percentile of daily precipitation and stream stage, respectively. The 98th percentile has been used as an approximation of the intense storms in a region. We hypothesized that streambank erosion would be greatest in epochs with frequent, intense precipitation because soils will have little time to dry between events. This follows Merritts et al. (2001) who observed that streambanks dessicate and fracture in summer months, which primes the banks for erosion that is most rapid in winter months due to freeze thaw. Although, we expect erosion along Richland Creek to be most rapid in summer months when precipitation events are most intense and where soils freeze less often than the soils in Pennsylvania.
Precipitation and stream gages were not present within the Richland Creek catchment during this study; although nearby gages provide sufficient data for this investigation. A rain gage (State Climate Office of North Carolina Station ID 317079) exists less than 6 km southeast of the study streambank (RG in Figure 1c). To determine if precipitation measured by this gage was representative of the area, we compared it to 3 nearby gages that encircle the streambank (USGS 02087182, 355020078465645, and 355856078492945). The maximum distance between these gages, and between a gage and the streambank is 24 and 18 km, respectively. The gages were compared using a correlation matrix of their daily precipitation totals when their operation coincided (28/09/11 to 27/06/12). The minimum correlation coefficient is 0.75 indicating that precipitation in this area of the Piedmont was similar and the gage designated as RG in this study can be used to infer precipitation at the streambank.

A stream gage was in operation downstream from the scanned bank along Richland Creek between 2002 and 2003 (RSG in Figure 1c). This gage was not commissioned during the 19-month monitoring period, although stream stage measured at this gage does demonstrate that Richland and Crabtree Creeks respond similarly to rain. The operation of this gage temporally overlaps a gage (CSG in Figure 1c) that is currently operating on Crabtree Creek (USGS Station ID 0208726005) about 400 m upstream from the confluence of these two streams. Five of the seven largest observed increases in stream stage coincide between Richland and Crabtree Creeks when the operation of gages overlaps (Figure 4). The magnitude of stream height response to precipitation is lower in Crabtree Creek relative to upstream drainage area. However, these streams respond similarly to precipitation meaning that the
Crabtree Creek gage is a reliable proxy for Richland Creek during the monitoring period.

**Figure 4.** Stream stage data of Richland and Crabtree Creek when the operation of these gages overlap: (a) Daily stream gage height normalized by the area upstream of the gage. (b) Dates (points) for which stream stage exceeded the 98th percentile stage of both creeks.

### 2.4.2. Field-based Experiment

Only the net streambank change integrated over several storms during an epoch of the monitoring timeframe can be analyzed because the bank was only scanned at the beginning and end of each epoch. For this reason, a field-based experiment was designed to simulate streambank wetting that might occur during a precipitation event. It was conducted on 1 August, 2012 after the 19-month monitoring period so
that experimental modification of the bank would not be recorded during the TLS monitoring surveys. Synchronous bank wetting, soil pore pressure measurements, and TLS scans were conducted every 20 minutes over a 4 hour, 20 minute period. 14 scans were collected during the experiment, resulting in 13 epochs that are henceforth referred to as Epochs 9 thru 21. Scans were acquired only from the location 12 m from the bank (scanner location 1 in Figure 1e) because the scanner could not be reassembled rapidly enough to allow multiple scan angles. Overhanging vegetation was removed during the monitoring period, so that line-of-sight obstructions were less of a concern for these scans.

The rate of bank wetting during the experiment was selected to simulate the most intense storm during the monitoring period, which occurred on 6 August, 2011. A hose was placed approximately 0.6 m from the edge of the streambank (Figure 5). Pressure-regulated drip emitters were installed along the hose every 0.5 m over 12.5 m, which extends 0.5 m past both ends of the scanned bank in order to eliminate edge effects upon the simulation. Water flowed through the hose at approximately 88 l/hr and wetted an area about 2.5 m². This is equivalent to a rainfall rate of 35 mm/h given the area wetted, which is approximately the rainfall rate of 28 mm/hr that occurred during the 6 August, 2011 storm. The antecedent hydrometeorological conditions of the ten days prior to this storm and experiment are summarized in Table 1. Precipitation during the ten days prior to this storm was slightly less frequent but more intense compared to the ten days prior to the experiment. Meaning, the amount of erosion that would result from an event similar to the 6 August, 2011 storm is likely under-predicted by the experiment because of drier
antecedent conditions and that bank subsurface was was only wetted along a narrow transect from above.

Figure 5. Field location set-up: (a) The TLS scanner was placed across from the streambank. The wetting transect was installed on the ground above the bank. Only scan location 1 in Figure 1e is shown. (b) Wire mesh representing the scanned bank in a transformed, localized coordinate frame. The grid cells are approximately 1 x 1 cm when projected on the x,z base plane. (c) Plan view schematic of wetting transect and tensiometers (T) with depths of 25, 55, and 85 cm. (d) A photograph of the experiment set-up taken from above the bank facing the TLS. Richland Creek crosses the center of the image. Inset shows a drip emitter along the wetting transect.
Pore pressure measurements and TLS scans were taken every 20 minutes while the streambank was being wetted. Soil pore pressure becomes less negative as soil saturation increases, such as would occur during rainfall. Spatiotemporal variability is highest in soils in-between rain events and decreases as the soil approaches saturation (Rezaur et al., 2002). To observe pore pressure variability as the subsurface approaches saturation, soil pore pressure was measured with six sets of tensiometers where each set included a tensiometer with a length of 25 (T25), 55 (T55), and 85 cm (T85). A correction factor was applied to the tensiometer reading to account for the weight of the water column within the tensiometer:

\[ T_s = T - (h \times 0.01 \frac{kp}{cm}) \]  

where \( T_s \) is soil water tension in kilopascals [kp], \( T \) is tension as measured by the tensiometer, and \( h \) is the height of the water column in the tensiometer in centimeters [cm] (Klute, 1986). Anomalous data caused by imprecise tensiometer septum punctures were removed prior to analysis. This occurred after Epoch 18 as the septum became worn from repeated punctures with the tensiometer needle. Anomalous readings were obvious when any of the following occurred: tearing of tensiometer septum, needle breakage, or atypical tensiometer values. During the experiment, 5% of the 234 readings were anomalous and so were not included in the analysis.
Table 1. Summary statistics of precipitation and stream stage for the antecedent conditions prior to two timeframes.

<table>
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<tr>
<th>Timeframe</th>
<th>Dates</th>
<th>Days</th>
<th>&gt; 0 [days]</th>
<th>Precipitation Mean ±1σ [mm]</th>
<th>Precipitation Mean recurrence interval [days]</th>
<th>Days above 98th percentile threshold*</th>
<th>Stage Mean ±1σ [m]</th>
<th>Days above 98th percentile threshold*</th>
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<td>Prior to 6 August, 2011 storm</td>
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<td>30%</td>
<td>35.8 ± 9.4</td>
<td>1.67</td>
<td>2</td>
<td>1.28 ± 0.32</td>
<td>2</td>
</tr>
<tr>
<td>Prior to field-based experiment</td>
<td>21/07/2012 to 31/07/2012</td>
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<td>1.71</td>
<td>0</td>
<td>1.39 ± 0.19</td>
<td>0</td>
</tr>
</tbody>
</table>

*The 98th percentile values of precipitation and stream stage during the monitoring period were 33.3 mm and 1.8 m, respectively.
2.4.3. TLS Data Acquisition and Processing

TLS is well suited for quantifying streambank erosion over sub-annual scales because it can be used to create highly detailed surface models (Resop and Hession, 2010; O’Neal and Pizzuto, 2011). The utility of TLS to measure erosion can be increased when it is combined with other techniques, such as when it is combined with analyses of aerial photographs (Day et al., 2013). In the present study, a field-based experiment was conducted to test hypotheses formed during TLS monitoring.

Methodologies to acquire, process, and analyze TLS data were identical in the 19-month monitoring period and the field-based experiment except for the use of a single, as opposed to two, scan location during the experiment. A Leica Geosystems ScanStation 2 lidar scanner was used to acquire $x,y,z$ coordinate datasets (“point clouds”) of the bank surface. The scanner is mounted on a static, leveled tripod and uses a rotating sensor-head coupled with an oscillating mirror that enables a 360 x 270° maximum field-of-view. With these capabilities, one scan location could capture the 11.5 m length of the bank, although thin vegetation obscures some of the bank surface. For this reason, two scanning locations for the 19-month monitoring period were used to minimize field-of-view obstructions and maximize sampling resolution of the bank face. Scanning locations were located on the floodplain of Richland Creek 12 and 20 m from the bank (Figure 1e). These distances are well within the 134 m maximum range of this scanner, at which point surface albedo is 0.18, the approximate reflection coefficient of vegetation and soil.
Data collected from the two scanning locations were co-registered (3-dimensionally aligned) to each other using six static targets within a localized coordinate system. Scan pairs of each survey were then co-registered to the scan pair acquired on the first date. The static targets were also used to co-register surveys collected on different dates because targets remained in place throughout the investigation. The coordinate system was defined by a 2D least-squares line that was fit to the \( x,z \) coordinates of lidar points collected in the first survey, similar to O'Neal and Pizzuto (2011). Meaning, the \( x \)-axis is the distance along the streambank, and the \( y \)-axis is the orthogonal distance between the bank face and a channel-oriented \( x,z \) vertical base plane (Figure 5b). The \( y \)-axis provides a relative length dimension from which changes in the bank surface can be measured.

Survey point clouds were filtered to remove predominantly vegetative non-surface points and then interpolated into DTMs using a regularized spline with tension (Mitasova et al., 2005). This methodology yielded 1 cm resolution DTMs of the bank surface for each survey where the raster cell value represents the orthogonal distance \( y \) between the bank face and the channel oriented base plane for that \( x,z \) location. Starek et al. (2013) provides more details on the data processing techniques used in this study.

2.4.4. Flux Characterization

Streambank evolution is characterized in terms of sediment flux in zones that coincide with the sedimentary layers. Flux among the zones was determined by segmenting the time-series DTMs by the height above the stream where
disconformities of the sedimentary layers exist. It is necessary to describe flux in term of zones since stratigraphic layers within DTM were only known by vertical position. Flux in lower zones is not only due to volumetric loss in the stratigraphic layer at that position, but also material transported into the zone from higher up the bank.

Sediment interlayer flux and export out of the scanned streambank area is analyzed as the changes in the streambank face (in the $y$-dimension) and volume. Flux was calculated by comparing the DTM created from data collected at the beginning and ends of each epoch, and is characterized as either a decrease (erosion) or increase (deposition) in the $y$-axis value or volume. The mean change of the bank face on the $y$-axis, $\Delta y$, along the length of the bank, $x$, was determined for each sedimentary layer, $l$, over each epoch, $t$, as follows:

$$\Delta y(i, t) = \frac{1}{n} \sum_{i=1}^{n} [y(i, l, t_e) - y(i, l, t_b)]$$

where $n$ is the total number of DTM raster cells along the stream length $x$ for a given sedimentary layer $l$, $t_b$ is the beginning survey and $t_e$ the end survey for epoch $t$, and $y(...)$ represents a DTM cell value. Erosion within a sedimentary layer $l$ at cell location $i$ during epoch $t$ is indicated when the quantity $[y(i, l, t_e) - y(i, l, t_b)] < 0$, whereas deposition occurred when $[y(i, l, t_e) - y(i, l, t_b)] > 0$.

The volume of material eroded during each epoch was quantified using 3-dimensional voxel models of the streambank. Voxel models represent terrain as volume elements called voxels that are analogous to grid cells in 2-D DTMs. In this application, the resolution of the voxel models was 1 cm, which was the same
resolution as the 2-D DTM. The volume of erosion during an epoch was determined by differencing voxel models constructed from scans acquired at the beginning and end of each epoch.

2.5. Results

2.5.1. Frequency of Intense Precipitation and High Stream Stage

The mean daily precipitation across the survey epochs remained similar, although the frequency and daily intensity varied (Figure 6a, 6b; Table 2). Precipitation during the gage correlation period (August 2002 to November 2003), the period of time when the operation of Richland and Crabtree Creek gages overlap, was more frequent than precipitation during the 19-month monitoring period: the percentage of days with precipitation was higher by 12% and the mean precipitation recurrence interval was 1.0 ± 1.9 days compared to 3 ± 3.7 days during the monitoring period. During the 19-month monitoring period, the wettest and driest timeframes were Epochs 7 and 4, respectively. Precipitation during Epochs 3 and 4 can be characterized as relatively infrequent with a mean recurrence interval that was 1 day longer than other epochs; although intense storms were most frequent during Epochs 3 and 4 (Table 2). The most intense precipitation during the 19-month monitoring period occurred during Epoch 3 when a storm delivered rain at a rate of 28 mm/h for three hours on 6 August, 2011. The 98th percentile threshold of 33 mm was exceeded on nine days during the monitoring period. These precipitation events were confined to seven months of the monitoring period: April to October in 2011 (Epochs 3 and 4), and May in 2012 (Epoch 7).
Figure 6. Gage data and streambank volume reduction during the 19-month monitoring period: (a) Daily precipitation total and maximum stage of Crabtree Creek. (b) Dates (points) for which stream stage exceeded the 98th percentile stage for Crabtree Creek and daily precipitation amounts from the nearby state climate office rain gage. (c) Volume of bank material removed during each epoch.
Table 2. Summary statistics of precipitation, stream stage, and volumetric change for the duration of the study.

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<th>Recurrence interval [mm]</th>
<th>98th percentile [mm]</th>
<th>Stage: Crabtree Creek Mean ±1σ [m]</th>
<th>98th percentile [m]</th>
<th>Volume change [m³]</th>
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<td>1.12 ± 0.22</td>
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</table>

*Richland Creek gage.
Stream stage of Crabtree Creek was similar throughout the gage correlation and 19-month monitoring periods. The daily mean was slightly higher in the former period, potentially reflecting the more frequent precipitation during this time. In the 19-month monitoring period, the 98^{th} percentile threshold of 1.8 m was exceeded 12 days during Epochs 3, 5, 6, and 7. The highest stream stages often reoccurred within a few days of each other because the stream more readily swelled above this threshold when the water level was already high.

### 2.5.2. 19-month Monitoring Period
Morphological changes of the streambank face throughout the duration of the monitoring period are apparent in TLS data. Change of the streambank face, \( \Delta y \), is largest along vertical columns in the mill pond zone, in the lower pre-European zone, and at the base of the bank (Figure 7). The most widespread surface change occurred in Epoch 3. This change is dominantly in the legacy sediment zones (early European and mill pond). Later loss is mostly in the pre-European zone at the base of the bank. Alternating loss and gain over epochs occurs along the upstream-most ~1.5 m of the bank. This section of the bank was highly vegetated and included a large tree that was precariously leaning over the edge of the bank. The pattern of observed surface change in this section could therefore be influenced by vegetation change not accounted for in the DTM time series, or the progressive effect of the leaning tree during the investigation.
**Figure 7.** Comparison of surface flux during the 19-month monitoring period and field-based experiment: (a) Shaded relief representation of the streambank from last epoch of respective timeframes. Stratigraphic layers are described in Figure 2. The streambank axes convention is presented in Figure 5b. Tensiometer locations are illustrated in the shaded relief image of the Field-based Experiment. Three sets of tensiometers are visible from the point-of-view of this figure, although 6 sets were used (see Figure 5c). (b) Change of the streambank face in the $y$-dimension of the stratigraphic layers, $\Delta y$, is defined in Equation 2. Red signifies net erosion and blue signifies net deposition along the $x$-dimension of the streambank. The alternating pattern of positive and negative change in the upstream-most ~1.5 m of the bank is adjacent to a leaning tree that is outside of the field of view of this figure.
Stratigraphic and temporal trends of erosion and deposition were inferred from the voxel representation of the bank. Erosion and deposition were calculated by summing volume change only at the cells where volume decreased and increased, respectively, during an epoch. These calculations were normalized by day because the lengths of epochs varied from 19 to 139 days. Net change was calculated as the cumulative difference between erosion and deposition of each epoch. The mean erosion, deposition, and net change across epochs for the entirety of the surveyed streambank were 0.002 ± 0.001 m$^3$/day, 0.024 ± 0.020 m$^3$/day, and -0.022 ± 0.020 m$^3$/day, respectively. These values demonstrate that the bank tended towards erosion. These values do not illustrate the daily changes as this is expected to have varied widely due to soil moisture conditions. Erosion was greatest during Epoch 3 when the volume of the bank was reduced by 8.4 m$^3$ (Figure 6c). This anomalously large amount of erosion cannot be explained by the longer time span of this 139-day epoch. Volume was reduced by only 6.2 m$^3$ during the other 424 days of the monitoring period. Epoch 3 coincides with the period of most frequent, intense precipitation during the monitoring period (Figure 6).

The amount of erosion per epoch of the stratigraphic layers is presented in Figure 8a. During each epoch, erosion is greatest in the legacy sediment zones (early European or millpond) with the exception of Epoch 4. The percentage of erosion that occurred during each epoch normalized by layer thickness is presented in Figure 8b. The greatest amount of erosion in the pre-European layer occurred in Epochs 2 and 4 following the large amounts of erosion in the Millpond layer in Epochs 1 and 3.
Figure 8. Estimated volume loss: (a) Total loss per epoch. Only cells with negative change were summed to determine the volume loss during each epoch and for each layer. (b) The values displayed in (a) are normalized by the thickness of the layers and the total volume loss per epoch.

2.5.3. Field-Based Experiment

The spatial variability of streambank flux was examined by simulating surface water infiltration that might result from an intense precipitation event. Much of the bank did not change outside of the magnitude of uncertainty during the experiment (Figure 7). The volume of water did not reach the amount necessary to develop a perched water table above the pre-European layer. However, bank change was detectable where preferential water flow paths were observed along vertical desiccation and horizontal seepage cracks (Figure 7). Soil pore pressure gradually
increased in all tensiometers (Figure 9). Pore pressure measured at the Set 5 tensiometers located behind the center of the streambank was much higher than the values at the other locations. This section of the streambank had a well-developed crack that ran parallel to the bank edge. This crack could have directed water to this section of the streambank, locally increasing pore pressure. This section approached saturation rapidly beginning at Epoch 17; although the effect of the bank-parallel crack on pore pressure was noticeable at the beginning of the experiment.

The experiment likely demonstrated a low estimate of flux that would result from a storm with the level of intensity simulated because (1) the simulated rainfall rate along the transect was 7 mm/hr greater than the most intense 1-hour storm recorded by the rain gage, and (2) the volume of precipitation reaching the forest floor would be even lower than recorded by a rain gage because the canopy would intercept a portion of the precipitation. Precipitation was simulated along a transect only about 0.2 m wide that was parallel to the length of the bank (Figure 5), so the total volume of water entering the subsurface was likely much more than the amount of water applied to the streambank during the 4-hour, 20-minute experiment. Additionally, elevated bank saturation prior to the experiment was not indicated by gages or field observations. The gage recorded 6 and 18 mm on 2 of the 3 days prior to the experiment. Days with precipitation between 6 and 18 mm occurred frequently during the monitoring period (Figure 6). The stream was at baseflow prior to the experiment, indicating that soil moisture was not at saturation levels. Decreased flux during the experiment would result from two-dimensional streambank wetting and low antecedent soil moisture.
Figure 9. Soil pore pressure recorded by tensiometers (T) with depths of 25, 55, and 85 cm during the Field-based Experiment (see Figure 5 for spatial distribution of tensiometers with respect to the streambank). The mean rate includes tensiometers from all depths.

2.6. Discussion

2.5.1. Boundary Conditions that Drive Streambank Erosion

The largest volume of sediment was removed from the mid-to-upper streambank. Gage data indicate that mechanisms other than fluvial processes are responsible for
the mid-to-upper bank erosional hotspot. The water level of Richland Creek appears to rarely rise to the base of the mill pond stratigraphic layer that resides 1.5 meters above the base of the streambank. The approximate maximum height of Richland Creek stage during the 19-month monitoring period can be inferred using the Richland and Crabtree gages when they were in operation at the same time. The range of stream stage at the Richland Creek gage was 0.57 m during the gage correlation period, which is nearly 1 m below the base of the mill pond legacy sediment unit. This range is likely similar or larger than what occurred during the monitoring period due to it being from a wetter period. Precipitation during the gage correlation period was higher: daily average of 11.0 ± 17.2 mm versus 3.4 ± 9.1 mm during the monitoring period. This indicates that the stream level of Richland Creek, as recorded by stream gages, did not reach the height of the mill pond layer along the bank. This follows our observations of the bank during TLS surveys. The high water mark was well below the base of the mill pond layer.

The notion that the Richland Creek stage has to reach the mid-to-upper bank to entrain large volumes of sediment and increase stream turbidity is further invalidated by data collected from automatic samplers as documented by Wegmann et al. (2013) for a storm event spanning February 4–5, 2011. In their analysis, stream stage increased about 17 cm before falling 5 cm between the 13th and 22nd hour of sampling, and then continued to increase past the final sampling hour (Figure 3; Wegmann et al., 2013). A return to base level turbidity values is observed during this lull in the stream stage upturn. The final two hours of sampling exhibit dramatic increases in turbidity concentrations at the downstream sample site, reaching over 3 times the turbidity at the upstream sampler. However, during the 3-fold increase in
turbidity the stream stage increased only 20 cm. This indicates that the swelling stream did not directly cause erosion of the higher bank material during this storm. Increases in turbidity are most likely attributed to fluvial transport of previously eroded sediment deposited at the base of the streambank. This is important because the stream attained NTU concentrations above local TMDL standards, even though (1) it is unlikely that newly detached bank material was entrained, and (2) the storm intensity represented in Figure 3 occurred about every 30 to 90 days during the monitoring period (Figure 6). These data demonstrate that even slight increases in stream stage—triggered by a storm intensity that is frequent—can cause stream turbidity to become elevated above water quality standards.

Significant streambank erosion occurs in response to large amounts of precipitation over extended wet periods following observations in prior studies (e.g. Simon et al., 2000). Flux of streambank material from the Richland Creek site was greatest in timeframes when intense precipitation was frequent. Much of the surface and volumetric flux during the 19-month monitoring period occurred in Epochs 3, 4, and 7 when intense, frequent storms occurred (Figures 6 and 7). Stream stage was also higher in these epochs, although significant flux in the lowest portion of the streambank did not always occur, indicating the importance of processes other than fluvial erosion for the delivery of sediment to the lower bank.

2.6.2. Processes that Deliver Legacy Sediment to Streams

Over timespans longer than this investigation, it is likely that the streambank face retreats in a parallel manner, removing the entirety of the meters thick sequence of
legacy sediment as the stream shifts along the valley bottom. Over annual or shorter
time scales, erosional mechanisms and volume removal varies over space and time.
Fine scale factors may act as an internal control on erosion susceptibility, such as
heterogeneities within stratigraphic layers and vegetation. The intent here is to
describe the physical processes that erode the streambank at the stratigraphic layer
scale and the dominant boundary condition (frequent, intense precipitation events).
While we did not scan the bank before and after singular events, field observations,
experiment results, and precipitation variability among epochs provide sufficient
indication of erosional process dominance for the Richland Creek streambank.

A suite of processes act to detach sediment from the streambank and entrain it into
Richland Creek. Mass wasting of sediment columns removed the greatest volumes
during the monitoring period. Mass wasting was confined to the approximately 2 m
thick mill pond stratigraphic layer during the 19-month monitoring period.
Processes other than stream undercutting must drive mass wasting given that
stream flow did not reach the base of the mill pond layer and the lower bank did not
exhibit recessed (concave-inwards) topography on any of the DTMs.

The field-based experiment was designed to test the hypothesis following from our
observations of the bank during the 19-month monitoring period: increased
subsurface porewater during, and subsequent to precipitation, drives mass wasting.
The volume of water to induce groundwater seepage erosion was not reached
during our experiment. However, the existence of preferential flow paths along
desiccation and seepage cracks that delineate the vertical and horizontal edges of
slumping sediment columns, respectively, were indicated by the coincidence of
areas with sediment column morphology, and water flow and sediment removal (Figure 7). The importance of preferential flow paths was also indicated at the location of tensiometer set 5, which was located behind a column of soil that appeared to be in the process of detaching. Soil pore pressure at this location was high at all depths throughout the experiment (Figure 9). On the ground surface above the bank, a deep (~1 m) vertical crack was observed near this set of tensiometers and behind this column. This section of the bank shares similarities with a bank in Warwickshire, United Kingdom described by Couper and Maddock (2001). Sediment columns of this silt-clay composed bank also detached along cracks perpendicular and parallel to the bank face. The columns were not removed at the time that desiccation cracks formed, but rather susceptibility to failure increased after the columns were formed by desiccation (Couper and Maddock, 2001). In the instance of Richland Creek, water flow that infiltrated the ground above the bank preferentially flowed into cracks that bounded sediment columns, which further increased mass wasting susceptibility of sediment columns.

The susceptibility of the mill pond layer to mass wasting-induced by shallow subsurface flow indicates that subaerial processes play an important role. The base of soil columns coincide with the mill pond-early European sedimentary unit contact where bank material transitions from silt-clay to sand. Streambank soils with high silt-clay content are typically susceptible to subaerial erosion that can then lead to increased mass wasting (Couper, 2003). The exposed streambank face along Richland Creek is dominantly columns of soil. Frequent, intense precipitation and the high summer air temperatures in this region encourage desiccation and
detachment of streambank soils. These factors suggest that desiccation is an important erosional process along this streambank.

The following process suite dominated streambank erosion and delivery of sediment to Richland Creek during this investigation. Soil columns detached along desiccation cracks and potentially were undercut where groundwater seepage occurred at the base of the early European layer. Mass wasting delivered large blocks of soil columns to the lower bank, or into the stream. This material was mostly cohesive legacy sediment with high silt-clay content that was less susceptible to fluvial entrainment, and as such acted to protect the streambank toe from rapid erosion (e.g., Simon et al., 2000; Couper, 2003). However, increases in bed shear stress along the bank toe, even during moderate precipitation events, were sufficient to entrain a portion of this sediment. This is made evident (1) by the high volume loss of MPLS in Epochs 1, 2, and 3 that likely resulted in increased deposition of these sediments at the toe of the bank, which was then preferentially entrained during Epochs 2, 4, 5, and 6 and is shown as greater erosion in the pre-European zone (Figure 8), (2) in epochs with low precipitation, such as Epoch 1, when net volumetric loss of the bank toe still occurred despite only infrequent, moderate precipitation, and (3) the automatic turbidity samplers documented by Wegmann et al. (2013) that recorded substantially increased turbidity following a moderate precipitation event (Figure 3).

Erosion was greatest in summer months along the Richland Creek streambank. This is contrary to other studies (e.g. Wolman, 1959; Hooke, 1979; Wynn et al., 2008) where streambank erosion was most severe in winter owing to more frequent freeze-thaw cycles; although process dominance at Richland Creek follows Green et
al. (1999) that described mass wasting-dominated streambank due to desiccation cracking. The volume of the streambank eroded in the summer would be decreased if desiccation was not an important process along Richland Creek, which suggests that historic land use can alter the seasonal primacy of erosion. Desiccation cracks were visible almost exclusively in the millpond layer and were the planes of weakness by which mass wasting occurred. Dessiccation-induced mass wasting would be less prevalent in millpond sediments in regions where wetting and drying cycles are less severe than the central Atlantic Piedmont where summers are characterized as hot with intense rain. To consider when erosion would dominant in the absence of legacy sediment, we could determine when erosion was most severe in the pre-European layer. However, it is not possible to consider this given that volumetric changes in the pre-European layer also include deposition from the legacy sediment layers. Future studies could quantify the impact of mill pond sediment on seasonal erosion primacy by quantifying streambank processes above and below breached historical mill dams. Studies could also examine the seasonal changes in local hydraulic conditions that may alter channel bedforms and shear stress at the bank toe. In the present study, analyses of channel bedforms and shear stress would likely not effect our interpretations of process dominance because fluvial erosion was minimal during the monitoring period.

Here we discuss the limitations of extending our results to other locations. The streambanks along the entirety of Richland Creek are likely eroding across orders of magnitude in terms of volume. Extrapolating the amount of erosion we measured along the 11.5 m reach to all of Richland Creek or other streams that incise into millpond sediment was not an objective of this research. Our intent was to
characterize the spatiotemporal patterns of erosion processes where they are active relative to hydrometeorological patterns. We expect that mass wasting is important along the length of the stream that is incising through millpond deposits because of the similarities in stratigraphy and morphology of the bank above the former milldam location. However, the importance of mass wasting decreases upstream as the height of the bank and thickness of the millpond layer decrease. Local climate and land use history should be considered when extending our results to other streams that are incising into millpond deposits. Merritts et al. (2011) describes streams in the northern Atlantic Piedmont that eroded by mass wasting of sediment blocks, although freeze-thaw was identified as an important process at this high latitude location.

This study demonstrates that landforms influenced by intensive historic land use practices can alter process dominance of streambank erosion. Historic periods of rapid upland soil erosion and stream impoundment lead to aggradation of thick packages of legacy sediment along Piedmont valley bottoms. These packages are well above local base level, especially where mill dams have been removed. Mass wasting, a process that is effective at adjusting hillslopes to a lowered base level at larger scales (Larsen and Montgomery, 2012), is also effective at returning streambanks to their pre-disturbance positions. Mass wasting would not be an important process along Richland Creek if this streambank did not contain the mill pond stratigraphic layer that is prone to desiccation, or the contrast in permeability between the early and pre-European layers that leads to groundwater seepage erosion. This study also supports the growing body of work that indicates that modern stream water quality is impacted by historic periods of rapid upland soil
erosion and stream impoundment (e.g., Jackson et al., 2005; Walter & Merrits, 2008; Pizzuto and O’Neal, 2009; Mukundan et al., 2010; Voli et al., 2013).

2.7. Conclusions

Analyses of stream gage data and TLS-derived DTMs of an 11.5 m long bank surface indicate that the bank is destabilized by desiccation-cracking where frequent, intense precipitation events induce slumping and the eventual collapse of sediment columns. Preferential flow paths along column edges remove grains and reduce soil column cohesion and tensional strength, which drives streambank collapse. Heightened stream stage during rain events entrains sediment along the toe that was previously detached from the bank. This material, much of it legacy sediment along Richland Creek and many other Piedmont streams, episodically increases stream turbidity during and after precipitation events.

Previously dammed streams will likely remain turbid for many years given the large stores of milldam deposits along valley bottoms. Stream restoration and turbidity reduction plans should take into account these potential sources for improved decision making and management strategies. Other, modern-day contributors of stream turbidity should be especially curtailed in these drainages because legacy sediment will be a persistent cause of turbidity even during moderate precipitation events.
2.8. References


Chapter 3

Modeling Streambed Median Grain Size to Locate Sediment Transport Nonlinearities

3.1. Introduction

The unparalleled diversity of freshwater mussels (Mollusca: Unionoida) in the southeastern United States is rapidly diminishing primarily due to habitat destruction and modification (Wiliams et al., 1993; Neves et al., 1997; Bogen, 2008). Increased upland and streambank erosion driven by intensive land use practices alter the geomorphic, chemical and biologic processes that have been associated with the endangerment and extinction of mussel species (Aldridge et al., 1987; Watters, 1999). Recently, attention has shifted to the importance of stream hydraulics upon the realization that mussel distribution has also been altered where habitat is not limited by poor water quality, limited food availability, impoundments, and loss of fish host species (Gangloff and Feminella, 2006).

Links among freshwater mussel distribution, stream hydraulics, and sediment supply and composition have been identified in previous investigations. Juvenile settling in streambeds, mussel burrowing ability, hyporeic water flow, and bed stability are requirements of productive mussel aggregations. In addition, sediment composition and grain size, and hydraulic shear stress are also important controlling environmental factors (Maio and Corkum, 1995; Brim-Box et al., 2002; Howard and Cuffey, 2003; Allen and Vaughn, 2010; Daraio et al., 2010). Models of reach-scale
stream hydraulics demonstrate that mussel abundance is lower where higher flows increase shear stress and destabilize the streambed (Gangloff and Feminella, 2006).

Streambed stability is governed by the competence of sediment grains on the bed, which is a function of the time integrated frequency and magnitude of discharge and sediment grain size (e.g. Buffington and Montgomery, 1999). Grains are incompetent, meaning that they have high potential to be entrained by stream flow, where the inertia of grains is overcome by the shear stress of water flow on the streambed, \( \tau_b \). The magnitude of shear stress that is required to entrain sediment grains is largely dependent on the critical Shields parameter, \( \tau^{*} \), that varies by grain size. The erosion mechanisms and lithology within the catchment are also an important control of streambed grain size; as sedimentary bedrock has the propensity to erode along interstitial grain boundaries (i.e. sandstone yields sand and conglomerate yields gravel). Predicting grain competence of mixed gravel-sand streams, a stream type common in the southeastern US, is complicated by hiding effects where fine grains are less susceptible to entrainment when they exist in a gravel framework and when fine grains interlock coarser grains (Sear, 1996; Snyder et al., 2008). Additionally, both field and flume studies demonstrate that gravel mobility increases when sand is present as part of the bed, and this becomes important when sand content is as low as 10–30% (Wilcock et al., 2001; Gran et al., 2006).

Downstream increases in sediment and water discharge generate stream-long trends in channel slope, width, depth, and streambed grain size. Grains fine downstream by abrasion and size-selective transport and deposition (Paola and Seal, 1995). In
alluvial rivers, fining increases gradually with the exception of across the gravel-to-sand transition (10 to 1 mm). This transition occurs over distances from 30 km in large rivers, to as short as a few hundred meters in small streams (Smith et al., 1996), and has been attributed to increased particle abrasion rates for fine gravel and coarse sand, and gaps in the sediment size distribution (Sambrook Smith and Ferguson, 1996; Wilcock, 1998, Knighton, 1999).

Quantifying trends of sediment grain size along the lengths of large rivers is prohibitively laborious in investigations that are strictly field-based. This limitation has been recently circumvented with numerical models that predict the competent streambed grain size, $D$, along large rivers (Buffington and Montgomery, 1999; Montgomery et al., 1999; Wilkins and Snyder, 2011; Snyder et al., 2012). $\tau_b$ and $D$ was predicted with parameters derived from digital elevation models (DEMs) and locally calibrated values obtained from limited field measurements. In this study, $\tau_b$ was predicted by assuming uniform and steady flow with the following relationship:

$$\tau_b = \rho g n^{3/5} Q_2^{3/5} w^{-3/5} S^{7/10}$$ (1)

where $\rho$ is density of water, $g$ is acceleration due to gravity, $n$ is Manning roughness coefficient, $Q_2$ is equivalent to discharge with a recurrence interval of 2 years, $w$ is channel width, and $S$ is channel slope (e.g. Snyder et al., 2003; Wilkins and Snyder, 2011). Grains size was predicted as:

$$D_m = \frac{\tau_b}{(\rho_s - \rho) g \tau_m^*}$$ (2)

where $\rho_s$ is density of sediment, $D_m$ is the $m$-percentile grain size, and $\tau_m^*$, critical shields shear stress that mobilizes $D_m$ (Shields, 1936; Snyder et al., 2012). Model
results can be displayed as a stream-long map of $D_m$ values, which represents the grain size that is competent at that stream reach for a discharge with a 2-year recurrence interval, which is approximately bankfull flow conditions (Wilkins and Snyder, 2011). Wilkins and Snyder (2011) used this modeling approach to explore the stream hydraulics that distribute suitable substrate for salmonid reed construction through a river, which was then used to better inform stream restoration efforts.

Herein, we apply a similar model to predict streambed grain size with mussel populations from the Sandhills region of the upper coastal plain in North Carolina. The relationship between freshwater mussel distribution and the river-long effects of non-linear channel hydraulics has yet to be examined despite the knowledge that bed shear stress is a control on mussel habitat and the gravel-sand transition. Also prior to this study, there was little information about the freshwater mussel populations and the distribution of suitable habitat in these river systems. From the survey results, we are able to characterize mussel density and diversity as a function of river long trends in substrate size while taking other habitat requirements into account. Water quality and temperature, food availability, and host fish populations are also requirements of healthy Unionidae populations. The focus of this paper is the stream-long distribution of freshwater mussels and streambed sediment. Distinguishing natural versus anthropogenic drivers of mussel habitat degradation is a necessary precursor to constructing effective conservation and restoration plans.
3.2. **Field Setting**

The Sandhills region spans from Georgia to North Carolina in the southeastern United States. It separates the crystalline bedrock of the Atlantic Piedmont to the west from the sedimentary bedrock of the Coastal Plain to the east (Figure 1). This study is comprised of three river catchments in the northern Sandhills region: Little River (area=1248 km²), Drowning Creek (840 km²), and Rockfish Creek (805 km²). The river systems are predominantly underlain by poorly consolidated Cretaceous sandstone and interbedded mudstone that erodes slowly, predominantly by soil forming processes that reduce the bedrock to sand-sized grains. Some upper reaches of Little River and Drowning Creek incise into moderately consolidated Triassic conglomerate composed of gravels cemented within a sandy matrix; upon weathering these conglomerates generate both sand and gravel-sized grains (Figure 1).
Figure 1. (a) The Sandhills region is located in the southeast United States crossing Georgia (GA), South Carolina (SC), and North Carolina (NC). The northern portion is drained by the Cape Fear River (CFR) and the Lumber River (LR), a tributary of the Pee Dee River (PDR). (b) Mussel surveys were spread throughout these river systems. Bedrock that can erode to gravel is in the headwaters of Drowning Creek and Little River. A USGS stream gage in each stream system was used in the sediment grain size model.
Large portions of the catchments that we investigated are occupied by the Fort Bragg and Camp Mackall military installations. Fort Bragg is the most populous U.S. Army installation in the U.S., spans 659 km², and covers 22% of the Little River and 40% of Rockfish Creek catchments. Camp Mackall occupies 33 km² and covers 4% of the Drowning Creek catchment. Prior to this study, there was little information about the status and location of freshwater bivalve populations and the distribution of suitable habitat for freshwater mussels on Fort Bragg. Many streambanks within Fort Bragg have been substantially altered during military training activity and land use, including numerous landing strips, heavy artillery impact zones, and a dense network of unpaved firebreak roads. These barren areas are susceptible to increased overland flow and erosion and are a potentially significant source of sandy sediment in streams. The bedrock underlying Fort Bragg and Camp Mackall does not contain gravel-bearing units, so stream sediment derived from erosion within the military installations will be dominated by sand.

3.3. Methods

3.3.1. Model Framework

A numerical model was used to locate the gravel-sand transition (GST) along study streams by predicting the median ($D_{50}$) streambed grain size. $D_{50}$ was predicted with inputs from a digital elevation model (DEM) and parameters calibrated with field data. Headwater portions of the study streams cannot be resolved with this DEM. For this reason, the model was limited to the trunk river of each study basin downstream of these minimum drainage area thresholds: Drowning Creek at 100 km², Little River at 200 km², and Rockfish Creek at 25 km². This section of the paper
first provides the framework in which stream-long grain size distribution was predicted. This is followed by a description of the methods used to acquire input values, followed by our modeling approach.

A stream-long grain size map was produced from the following steps: (1) the shear stress acting on the streambed was predicted using Equation 1, (2) the critical Shields shear stress value at which $D_{50}$ can be mobilized in field-survey reaches was predicted using a rearranged version of Equation 2:

$$\tau_m^* = \frac{\tau_b}{(\rho_s-\rho)g D_{50}} \tag{3}$$

and (3) $D_{50}$ was predicted along the length of a stream $\tau_m^*$ from the following relationship:

$$D_m = \frac{\tau_b}{(\rho_s-\rho)g \tau_m} \tag{4}$$

where $\rho_s$ is 2650 kg/m$^3$, the density of quartz and feldspar, which comprises $>95\%$ of the channel sediment by volume.

### 3.3.2. Model Parameters and Variables

Critical shear stress values were approximated using sediment samples excavated from the beds of study streams. Sediment was collected as grab samples instead of the more often used pebble count method where a subsample of streambed sediment is measured grain-by-grain in the field. We choose not to do pebble counts because this method tends to under sample fine sediment, including sand, which is prevalent in the streams of this investigation. Instead we collected approximately 2 kg of sediment at each site by taking one 12 oz-scoop of sediment every 0.5 m for 5 m down the channel thalweg. Streambeds were sampled in this manner two to three
times at equal distances along study reaches. Samples were dried and sieved in the lab in six grain size fractions (0.5, 0.85, 2, 4, and 8 mm). Sieve stacks were mounted on a shaker that operated for 30 minutes and then size fractions were weighed. The two grain size fractions that bound $D_{50}$ were linearly regressed to estimate the median grain size to the nearest millimeter. Samples were separated into groups of gravel ($D_{50}$>2 mm) and sand ($D_{50}$<2 mm). The critical shear stress for gravel, $\tau_g^*$, and sand, $\tau_s^*$, sample groups were calculated using Equation 3 separately for each stream, which provided the range of critical shear stress values that was assumed to be representative of the entirety of the modeled stream length.

Topographic input values were extracted from a digital elevation model downloaded from the North Carolina Floodplain Mapping Program (2012). This DEM was created from lidar data collected in 2001 that has a horizontal resolution of 6 m and a vertical accuracy equal or less than 25 cm RMSE. The stream network was determined using the DEM, and flow routing algorithms in ArcGIS. Channel modification by human development and inaccuracies of lidar data caused channel centerlines to be incorrectly positioned in the modeled landscape. Stream centerlines were corrected where the ArcGIS-determined centerlines did not align with the stream channel visible in shaded-relief maps produced from the lidar DEM and aerial photographs. Correcting centerlines was a necessary step because improperly mapped centerlines would yield incorrect channel slope values. Accurately identifying channel slope is a vital component of these types of models given that this parameter ranges over several orders of magnitudes in many rivers (Snyder et al., 2012).
Channel slope was determined by following procedures outlined in Wobus et al. (2006) and Snyder et al. (2012) that addressed issues with sampling elevation data along channels. Elevation along the corrected centerlines was extracted from the DEM every 6 m, matching the resolution of the DEM. To ensure monotonically decreasing elevation downstream, profiles were smoothed every 25 points, and local minima were determined and then spline interpolated. The calculation of channel slope is sensitive to the length scale over which it is calculated. A sensitivity analysis of the selection of the length on the calculation of slope was conducted for all streams using 18, 102, and 1002 m reaches (multiples of 6 m, the resolution of the DEM) using central-differencing moving windows.

The model incorporated the range of Manning’s coefficient, $n$, values between 0.05–0.07, which is the range of values in engineering reports of the study rivers (North Carolina Floodplain Mapping Program, 2012). To estimate the 2-year discharge along the study streams, $Q_2$ was assumed to vary linearly with drainage area, $A$ (Snyder et al., 2003). The ratio of discharge and drainage area was determined for each stream system. The 2-year discharge at each stream gauge ($Q_{2\text{gauge}}$) in the three river systems is documented in Weaver et al. (2009). This value was divided by the drainage area upstream from the stream gauges. $Q_2$ is estimated at each cell along study streams by:

$$Q_2 = A(Q_{2\text{gauge}}/A_{\text{gauge}})$$  \hspace{1cm} (5)

Channel width was measured using a 2010 aerial orthophotograph with a horizontal resolution of 0.5 m. Vegetation overhanging streams obscured the banks of tributaries and the upper reaches of the three rivers. To determine channel width, $w$,
in these smaller stream reaches, the relationship between drainage area and channel width was determined along reaches in the lower sections of the three rivers, and extrapolated to upper reaches using the relationship:

\[ w = k_w A^b \]  \hspace{1cm} (6)

where \( k_w \) and \( b \) are empirical values. Channel width was quantified continuously at the scale of the DEM resolution along the three rivers where banks were visible. A mask of the channel was manually digitized for each river using the aerial photograph. Width was measured automatically using the RivWidth ITT Visual Information System IDL utility (Pavelsky and Smith, 2008).

3.3.3. Modeling Approach

Our modeling approach incorporated the uncertainty of model parameters and inputs in order to identify the stream-long distance over which the gravel-sand transition occurs. At each DEM cell along the study streams, \( D_{50} \) predictions were maximized and minimized given the ranges of critical shear stress by grain size, channel slope length scale, and Manning’s \( n \) values. First we determined the range of \( \tau_m^* \) for each study stream where the minimum was \( \tau_g^* \) and the maximum was \( \tau_s^* \). The calculation of \( \tau_g^* \) determined from the gravel portion of sampled sediment was further minimized by using a \( n \)-value of 0.07 in Equation 3 and adding the RMSE of channel slope calculated in 18 m versus 102 m central differencing windows to channel slope determined at the cell using the former window size. Conversely, the calculation of \( \tau_s^* \) was further maximized by using a \( n \)-value of 0.05 and subtracting the length scale RMSE from channel slope. Next, the minimum and maximum \( \tau_m^* \) values were used in Equation 4 to predict \( D_{50\text{max}} \) and \( D_{50\text{min}} \) respectively, and again
appropriate values of $n$ and channel slope inputs were selected to maximize the predicted range of $D_{50}$ at each cell. For both $D_{50 \text{max}}$ and $D_{50 \text{min}}$, the point where size declines below 2 mm is defined as the GST for that calculation. Meaning, the actual gravel-sand transition must lie in between the GST locations predicted by $D_{50 \text{max}}$ and $D_{50 \text{min}}$ values.

3.4. Results

3.4.1. Grain size model

Channel slope was calculated over 18, 102, and 1002 m reaches using a central-differencing moving window. Slope calculated over these lengths were similar (Figure 2). The 1002 m length scale was the least correlated to the 3-point, 18 m length scale especially at gentler slopes. The effect of the channel slope length scale in sediment grain transport is likely limited to the slope in the immediate vicinity of the grain, and at most, in the 100 m reach around the grain. For this reason and the fidelity of slope calculations over 18 and 102 m windows, channel slope was calculated over 18 m windows in the model.
Figure 2. Comparison of length scale calculations on channel slope. Slope calculated in moving windows that were 18 and 102 m long were nearly identical. Slopes calculated in 1002 m moving window differed from the 18 m-calculated slopes, especially where channels were nearly flat.

Measured and predicted grain size was scattered, although the model did correctly predict grain size in sand and gravel fractions with the exception of a single sample from Drowning Creek (Figure 3). The samples that were under-predicted by about 5 mm were collected from a reach of the Little River that was partially rerouted around a breached dam. The remaining dam structure increased channel slope, resulting in a coarser bed; although at the resolution of the DEM the channel was more similar to its pre-disturbance geometry, resulting in the inaccurate prediction of finer sediment.
Figure 3. Comparison of observed and predicted $D_{50}$. Gravel and sand fractions are separated at a threshold of 2 mm (dotted line).

3.4.2. Stream longitudinal trends in D50

Predicted grain size gradually decreases except at tributaries junctions and reservoirs (Figure 4). Model variables, channel width and stream discharge, were calculated from drainage area, which operates as a step function due to tributaries inputs. Reservoirs, basin headwaters, and other areas where the stream centerline could not be distinguished were not modeled, although channel modification up and downstream of reservoirs appears to have impacted predictions, notably along Rockfish Creek at approximately 5 km down the channel.
**Figure 4.** Longitudinal profiles and predicted $D_{50}$ of study streams. Smoothed long profiles (black line) originate at the drainage area threshold indicated in the text and are shown only where grain size was predicted. Non-smoothed profiles (blue line) originate at a drainage area threshold of 10 km$^2$ and are shown here only for reference. Both the upstream and downstream of limits of $D_{50}$ predictions (black and gray points) are shown. The location where the predicted $D_{50}$ crosses the 2 mm threshold of sand was defined at the gravel-sand transition. Locations of mussel surveys along these streams are denoted on the long profiles.
The gravel-sand transition (GST) of the maximum and minimum $D_{50}$ predictions varied by stream. The basin area limits of the GST along Little River and Drowning Creek were 377–677 and 344–582 km$^2$, respectively. Along Rockfish Creek, predicted $D_{50}$ crossed below the 2 mm threshold only at the uppermost end of the modeled portion of this stream; however, significant lengths of natural gravel patches were not observed in this stream, consistent with the lack of gravel-producing outcrops in the Rockfish Creek catchment (Figure 1).

3.5. Discussion

It was necessary in our investigation to take basin lithology into account when calibrating the critical Shields stress parameter with sediment collected from local streambeds. The bedrock of the Drowning Creek and Little River basins supply streams with larger grain sizes than what is available to streams in the Rockfish Creek basin. Our approach to identify the gravel-sand transition would not work if a universal critical Shields stress parameter was used, such as the commonly used value of 0.04 for gravel. Using this value would have resulted in large over-predictions of median grain size along Rockfish Creek.

The calibration samples for $\tau_g^*$ of Rockfish Creek were finer than gravels of other streams, which effectively provided model control upon grain size availability. These samples were collected from short gravel-dominated reaches where evidence of tanks fording the stream was apparent. The stream substrate at these tank crossing sites appear to have been augmented with non-locally derived gravels (conglomerate), likely imported for this purpose from outside the Rockfish Creek
basin. Either these constructed tank crossings, or small unmapped outcroppings of gravel-producing bedrock, supply Rockfish Creek with minor amounts of gravel that are not sufficient to produce the grain size trends observed in the other two study streams. Alluvial streams adjust their form to the imposed sediment supply, although the variability of supply magnitude and grain size composition, and the precision of the DEM in this investigation were not sufficient to rely on universal $\tau_m^*$ values and channel form to recognize the variability of the gravel-to-sand transitions across study basins.

The gravel-to-sand transition was observed at similar positions along Drowning Creek and Little River. These basins have similar lithology and size, although the amplitude of channel slope variation over sub-km reaches is lower along the Drowning Creek study reach, which contributed to a steadier decline in predicted grain size compared to the Little River. The upper prediction of $D_{50}$ along Little River straddles the 2 mm threshold for approximately 30 km. The streambed of this length of stream is especially susceptible to changes between gravel and sand dominance. Increased sand supply often leads to increased transport capacity of both gravel and sand fractions (Jackson and Beschta, 1984; Curran, 2007) that can lead to a lengthening of the gravel-sand transition (Knighton, 1999; Gran et al., 2006) and a reduction in optimal (stable) mussel habitat.

This model can benefit stream restoration efforts by itself, or coupled with upland erosion models. A stream experiencing a large volume of fine sediment input can be modeled so that reaches that contain the geometry for sustaining gravel substrate can be identified. This ideally would take place in conjunction with upland erosion
mitigation efforts to reduce stream sedimentation. Model failures can also be used to identify stream reaches for further investigation and restoration. The largest under-prediction of field samples was collected from a reach where a partially breached dam altered the channel and streamflow.

3.6. References


