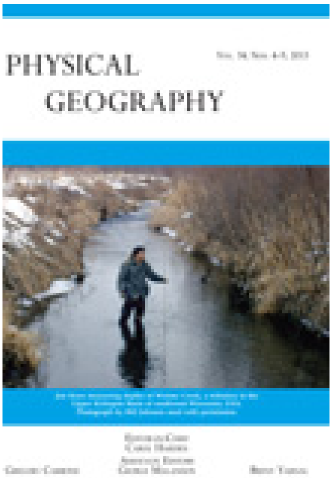

Physical Geography
Publication details, including instructions for authors and subscription information:
http://www.tandfonline.com/loi/tphy20

Modeling stream-bank erosion in the Southern Blue Ridge Mountains

James C. Rogers[a] & David S. Leigh[a]
[a] Department of Geography, The University of Georgia, Athens, GA, USA
Published online: 31 Oct 2013.

To cite this article: James C. Rogers & David S. Leigh (2013) Modeling stream-bank erosion in the Southern Blue Ridge Mountains, Physical Geography, 34:4-5, 354-372

To link to this article: http://dx.doi.org/10.1080/02723646.2013.846745

Physical Geography, 2013
Vol. 34, Nos. 4–5, 354–372, http://dx.doi.org/10.1080/02723646.2013.846745

Modeling stream-bank erosion in the Southern Blue Ridge Mountains

James C. Rogers* and David S. Leigh

Department of Geography, The University of Georgia, Athens, GA, USA

(*Received 12 March 2013; accepted 11 September 2013*)

Deforestation, followed by soil erosion and subsequent deposition of alluvium in valleys, played a critical role in the formation of historical terraces in much of the Southern Blue Ridge Mountains. Such terraces add a significant amount of sediment to the tributaries of the region as streams laterally erode the terrace banks. This study examined the contribution of total sediment yield derived solely from eroded stream banks in small watersheds (<20 km^2), using floodplain widths as proxies for long-term lateral erosion rates. The raw data were derived from watersheds with different land covers (Coweeta Creek and Skeenah Creek watersheds in the Upper Little Tennessee River basin). Bank-derived sediment yield estimates were modeled in a Geographic Information System, using linear regression to relate floodplain widths and erodible terrace bank heights. We found total stream length to be a good predictor of both lateral erosion rates and erodible bank heights. Land cover, basin/network morphometrics, and reach-scale stream conditions were not good predictors. Modeled lateral migration and sediment yield results compare favorably to empirical measurements from five independent watersheds in the region. Modeled estimates fall within ±50% or better of the observed values, at 16.33 to 25.02 t km^{-2} yr^{-1}.

Keywords: stream lateral migration; sediment yield; watershed; morphometry; stream morphology; North Carolina

Introduction

Stream-bank erosion potentially adds a significant percentage of sediment to overall sediment yield in forested drainage basins (Knighton, 1998; Meade, Yuzyk, & Day, 1990; Nanson & Hickin, 1986; Reid, 1993; Reid & Dunne, 1996; Walling & Fang, 2003). In the southern Appalachians, a large amount of sediment was eroded in response to late 19th and early 20th century timber harvests and is stored within historical terraces in the smaller tributaries (Leigh, 2010). Currently, these deposits act as an important sediment source for locations downstream and within the associated watershed (Harden, 2004; Leigh, 2010). It is well known that large sediment inputs from relatively discrete events (i.e., timber harvests) in watersheds have lag and residence times that are critical to fluvial processes and geomorphic form (Kelsey 1982; Madej & Ozaki, 1996; Montgomery, 1999; Swank, Vose, & Elliott, 2001). Lateral erosion by large rivers has been extensively studied (Harden, Foster, Morris, Chartrand, & Henry, 2009; Hooke, 1979, 1980; Lawler, 1993; Lawler, Grove, Couperthwaite, & Leeks, 1999; Murgatroyd & Ternan, 1983; Simon, Curini, Darby, & Langendoen,

*Corresponding author. Email: jamescolinrogers@gmail.com

2000; Thorne, 1982; Wolman, 1959). However, the lateral erosion of smaller streams and their contribution to sediment yield is not as well understood or well documented in the Blue Ridge Mountains. This is especially true in the context of past and current land cover, yet understanding sediment movement is a critical component in stream management, rehabilitation, and restoration (Harden et al., 2009). This study models total sediment yield from banks within drainage basins smaller than 20 km^2 using measured width of the geomorphic floodplain as a proxy for long-term lateral migration rates. The model uses linear regression to estimate how lateral migration rates result in net sediment yield from terraced stream banks. We also consider other physiographic and land cover variables that may affect the system.

Study area

The sample sites fall within the southern portion of the Blue Ridge physiographic province in the watersheds of Skeenah and Coweeta Creeks of Macon County, North Carolina. These tributaries flow directly into the Upper Little Tennessee River, which flows northward and drains portions of northeast Georgia and western North Carolina (Figure 1). While both the Skeenah and Coweeta watersheds were extensively logged in the late 19th and early 20th centuries, land containing the Coweeta watershed was purchased in 1918 by the US Forest Service for the purposes of conservation and forest management. Today the Coweeta Creek basin represents a mostly forested basin (97% forest cover), while Skeenah represents a basin with housing development and small farms (73% forest cover). The forested sections generally are limited to the upland periphery in the Skeenah Creek basin, in which the USDA Forest Service prohibits development. The Coweeta Creek basin, like the Skeenah Creek basin, underwent vast forest clearance in the late 1800s and early 1900s; however, most of the basin is now forested and used for scientific research. The close proximity of these basins facilitates comparative study of ongoing landscape change since the time of timber harvest (Price & Leigh, 2006a, 2006b), which allows our study to consider a range of land-cover variables that could potentially affect lateral migration rates (Table 1).

During the turn of the 20th century, extensive harvesting of old growth forests for timber peaked in the Southern Appalachian Mountains of the USA (Ayers & Ashe, 1904; Eller, 1982; Glenn, 1911; Yarnell, 1998). The forests of the entire region had been completely harvested by the 1940s (Yarnell, 1998). Change in land cover disturbed many streams, some of which are still in the process of recovery (Leigh, 2010). People also were affected by disturbed landscapes via increases in the size, duration, and frequency of flooding. Stream valleys experienced much sedimentation due to the erosive nature of timber harvesting and farming on mountain slopes (Glenn, 1911; Leigh, 2010). Today, the legacy of past landscape erosion is apparent in the hydrologic system, the sedimentological structure, and the geomorphologic characteristics of floodplains and terraces in the Southern Blue Ridge (Leigh, 2010; Leigh & Rogers, 2007; Leigh & Webb, 2006; Price & Leigh, 2006a).

Deciduous hardwood forests represent the dominant land cover in the study area, both currently and throughout the Holocene (Roosevelt, 1902; Yarnell, 1998), and small amounts of other land-cover types comprise the remainder. The bedrock in the study area is primarily biotite gneiss and quartz dioritic gneiss (Robinson, Lesure, Marlowe, Foley, & Clark, 1992). The 30-year (1981–2010) average precipitation at the Coweeta Experiment Station's low elevation station (at 685.5 m above sea level) is 1752.3 mm per year, with a monthly high of 170.4 mm during the month of February, and the average 30-year

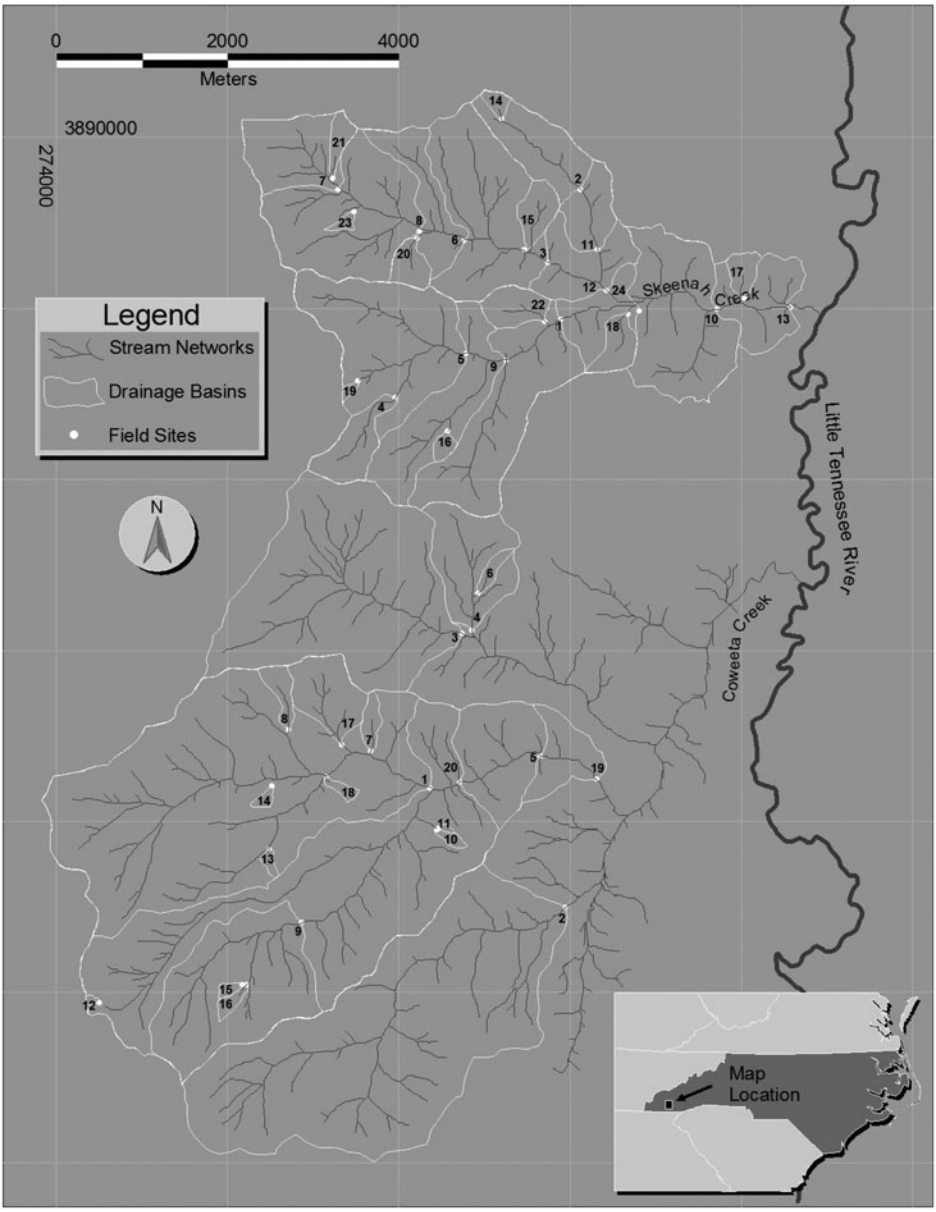

Figure 1. The study area, including sampled stream reach locations and the associated reach's drainage basins.

annual temperature is 13 °C. Average 30-year monthly temperatures for January and July are 2.5 and 22.5 °C, respectively (National Climatic Data Center, 2011).

Methods

Data were collected from August 2005 to May 2007. Sampled steam reaches were selected in a pilot study using Jenks Natural Breaks Classification Method as a

Table 1. Coweeta and Skeenah drainage basin characteristics.

	Coweeta	Skeenah
Drainage area (km^2)	40	18
Perimeter (km)	29	22
Maximum elevation (masl)	1591	1085
Minimum elevation (masl)	640	622
Total relief (m)	951	463
Average slope (%)	47	26
Forested land cover (%)	95	73
Impervious surfaces (%)	0.21	0.55

stratified random sampling scheme based on Shreve stream ordering system within each basin (Shreve, 1967), as explained in Rogers (2011). This scheme resulted in a diverse representation of land-cover types, physiographies, stream sizes, and morphological conditions. The stream network was delineated using a 10-m horizontal resolution Digital Elevation Model obtained from the US Geological Survey and a stream-initiation threshold of 400 pixels (4 ha basin area) in the ArcView 3.3 (2003) extension "Basin 1" (Petras, 2003), which objectively delineates stream networks. To ensure discharge and morphological consistency, sample reaches were selected at least 5 m downstream and 5 m upstream from anthropogenic structures and did not intersect any other stream, pipe, or culvert along the surveyed reach.

Stream channel data were collected only during baseflow conditions. A total of 41 stream reaches were sampled, including three reaches derived from an independent and compatible study (Price & Leigh, 2006a, 2006b). To ensure complete representation of the reach, the surveyed reach length was 30 times the average wetted width of the stream (Simon & Castro, 2003). Basic cross-sectional data were collected at distance intervals of two times the wetted width along 16 transects perpendicular to the direction of flow.

The primary data (ultimately used as dependent variables in our models) collected at each of the 16 transects included the geomorphic floodplain width (if present, measured horizontally from top of the channel bank to the base of the terrace or hillslope scarp) bank height (measured vertically from the thalweg to the top of the bank), and designation of whether the bank was composed of floodplain, terrace, or hillslope materials. Floodplain width represents the minimum lateral distance the stream has traveled since floodplain initiation and, as such, provides a minimum estimate of long-term lateral erosion rates (i.e., lateral erosion rate = floodplain width/years since floodplain initiation). It is important to note that the estimate of lateral migration distance provides a minimum rate, because it assumes unidirectional migration, which is not always the case. Despite the assumption of unidirectional migration, this is adequate to evaluate net sediment yield from stream banks that are terraces, because net yield from lateral erosion into floodplain deposits is likely to be zero (new floodplain deposits counterbalance floodplain erosion). We recognize that other sources of sediment are present, but our intention is solely to isolate centennial-scale average values of bank erosion into terrace remnants. Based on luminescence dates and historical evidence, Leigh (2010) established that the modern floodplain formed ca. A.D. 1915 (± 21 years, or the midpoint between 1894 and 1936 A.D.), following incision beneath an historical terrace. Thus, we assume A.D. 1915 ± 21 years provides the time-zero for lateral erosion estimates from floodplain widths. This equates to 91 years for our study, with a range from to 112 to 70 years. Bank heights were used to estimate the erodible bank height above the floodplain elevation. This erodible bank height is defined as the reach-averaged observed height of the terrace/hillslope stream bank minus the reach-averaged height of the floodplain above the channel bed. Multiplying this erodible terrace bank height by the length of the geographic information system (GIS)-derived stream segment between confluences (nodes) in the stream network provides an estimate of the total area of the eroded bank, within an individual stream segment, that exceeds the amount of sediment replaced on the floodplain. Combined with the estimates of lateral erosion rates, this allows a net bank-sediment yield for the stream segment to be derived.

Additional data used as independent variables in regression models to characterize the slope, cross-sectional, and sedimentological conditions of the stream were measured using a combination of standard stream assessment methods. These methods are described by US federal agencies, including but not limited to the US Geological Survey (Fitzpatrick et al., 1904), the US Department of Agriculture (Harrelson, Rawlins, & Potyondy, 1994), and the US Environmental Protection Agency (Kaufmann, Levine, Peck, Robison, & Seeliger, 1999). A full description of sample-collection methods is provided by Rogers (2011).

To explore relationships between stream lateral erosion rates and basin-wide physiographic and land-cover characteristics, spatial data were sampled for the respective drainage basins within groups consisting of morphometric, stream network, or land-cover variables. The GIS ESRI's ArcView 3.3® (2003) was used to generate, collect, and study these drainage basin characteristics upstream from the 0x point, or the most downstream surveyed cross-sectional transect at each study reach, which was the basin outlet point. The respective basin's area, perimeter, drainage network, and outlet elevation were collected using ArcView in conjunction with Basin1 (Petras, 2003). From this data-set, 19 drainage basin morphometric and 13 stream network variables were calculated and tabulated using Microsoft Excel (Rogers, 2011). A varying set of 16 land-cover characteristics were also collected (Rogers, 2011) for each basin using land-cover data from the National Land Cover Database (Homer et al., 2007).

Descriptive statistics were calculated using Microsoft Excel 2003 for each of the reach and basin variables per site, as well as between sites. Inferential statistics and normality testing were performed using SigmaStat (1997). Proportional (percent) data were transformed using the arcsine-square-root function. A value of one was added to all of the observations that contained zero values for the sake of sufficiently testing the normality of the data-set.

The confidence interval of Spearman rank-order correlation coefficient was used to eliminate independent variables that were not correlated to the dependent variables (floodplain width and erodible terrace bank height) at P-values greater than 0.05. The Kolmogorov-Smirnov goodness-of-fit test was used to test for normality and the Pearson product-moment correlation coefficient was calculated for normal variables. Variables with correlation coefficients (r-values) > 0.80 for erodible terrace bank heights and >0.60 for floodplain widths were accepted as adequate for predicting each respective dependent variable and constituted our final data set for exploring linear and multiple linear regressions as models for predicting lateral erosion rates. Similarly, erodible terrace-bank heights were also predicted using linear regression techniques that relied on the independent variables of basin topography, network size and character, basin morphometrics, reach slope, and land cover. We started with simple bivariate linear regression models and then used forward and backward stepwise regression to identify multiple regression models that could better predict floodplain width and erodible bank height. The stepwise multiple regression models eliminated independent variables that covaried with each other at r values > 0.8 from consideration in the same model.

Our final model of sediment yield from the entire watershed combined our linear regression estimates of erodible bank height and bank-erosion rates (derived from floodplain widths as width/years) into a GIS of the entire stream network. That is, the GIS identified the nodes of stream confluences, which delimited individual stream segments within the network, and our linear regression estimates were multiplied by the measured length of each individual segment to derive a volume of sediment yield from each segment (e.g., stream segment sediment yield = measured segment length × regressed erodible bank height × regressed erosion rate). Sediment yield from all individual stream segments were then summed for the entire tributary network. These volumetric amounts were then converted to a mass based on an average soil density of 1.3 g cm^{-3}, as observed by Price, Jackson, and Parker (2010).

To validate our sediment yield model, comparable sediment yield observations were sought from published sources in the region, and to isolate sediment inputs from banks (i.e., eliminating hillslope sources from agriculture, construction, logging, and urbanizing influences), these comparable drainage basins needed to be almost completely forested. In fully forested basins, stream channels (banks and beds) contribute the majority of sediment yield, although other, minor sources include tree throws, mass wasting, and animal burrowing. We identified five fully forested watershed sediment yield studies nearby, including three by Simmons (1993; Bee Tree Creek, Cataloochee Creek, and Nantahala River) and two by Royall (2000, 2003; watersheds of Deer and Thompson Lakes).

Results

The dependent variables used in the regression models were the geomorphic floodplain width and the erodible terrace-bank height (Figures 2 and 3). A total of 75 independent variables (shown in Rogers, 2011) were collected in the field or generated using a GIS to test for significant predictors of those dependent variables. The independent variables represented four general categories, including basin morphometry, basin stream network, basin land cover, and morphology/sedimentology of the stream reach. A process of elimination was used to pare down the independent variables to a set of variables that best predicted the dependent variables of erodible terrace-bank height and floodplain width.

Best predictors of erodible bank height and floodplain width

A first round of variable elimination used the Spearman rank-order correlation with a confidence interval (p value) of < 0.05 as the threshold for accepting the independent variable. Of the 75 independent variables, 39 were significantly correlated with floodplain width and 51 with erodible terrace-bank height. The Kolmogorov–Smirnov test was used next with a confidence interval threshold of $p > 0.05$ to eliminate variables that were not normally distributed. As they pertain to erodible terrace-bank height, only 20 of the 51 independent variables that passed the Spearman rank-order test have normal distributions. Of the 39 variables correlated with floodplain width, only 19 have normal distributions. The dependent variable of floodplain width was best normalized using a $\log_2$ transformation.

Figure 2. Geomorphic floodplain (in front of horse on left side) and historical terrace (on right side of the stream).

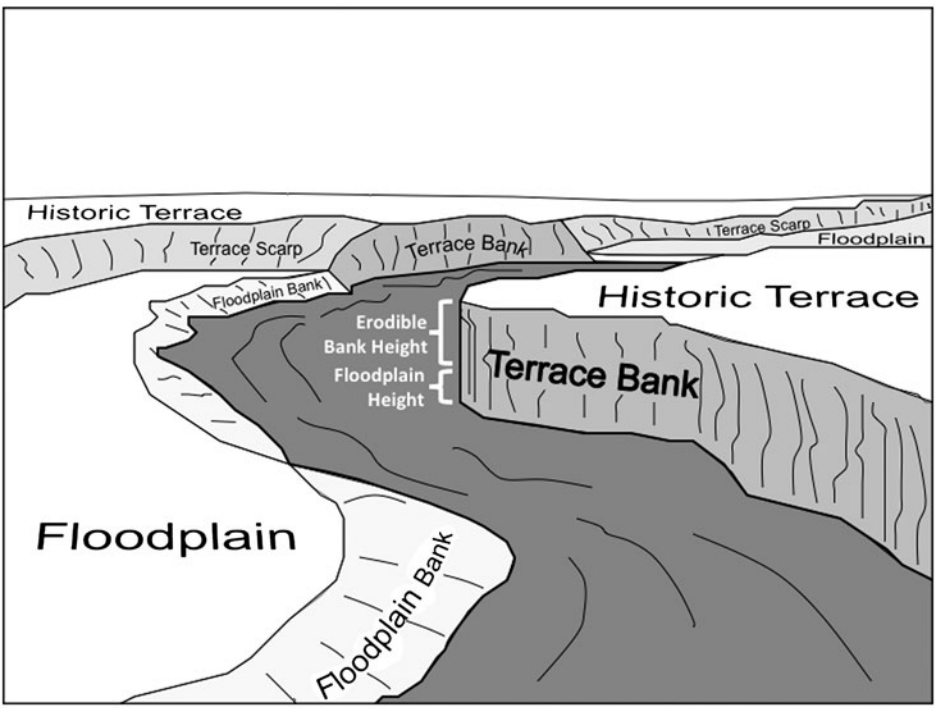

Figure 3. Schematic diagram of the geomorphic floodplain and the higher surface of the terrace that includes the “erodible bank height” above the level of the floodplain.

Table 2. Best bivariate correlates to erodible terrace bank height and floodplain width. All correlates have a significant value (*p*-value) < 0.0001.

Basin characteristic variable	Transform used	Correlation with terrace bank height (m)	Correlation with $\log_2$ of floodplain width (m)
Drainage area (m^2)	Log_{10}	0.83	0.61
Basin perimeter (m)	ln	0.84	0.6
Stream length longest path of segments (m)	ln	0.83	0.61
Stream segment lengths total (m)	Log_{10}	0.84	0.61
Stream segment length maximum (m)	None	0.84	0.63
Basin relief ratio	Log_{10}	na	−0.62
Relative relief (m/km)	Log_{10}	na	−0.62
Basin length (km)	Log_{10}	0.83	0.61
Basin length (m)	Log_{10}	0.83	0.61
Segment length thread average (m)	Square root	0.82	na

The Pearson product–moment test was used to find the strongest correlations for both of the dependent variables – floodplain width and erodible terrace-bank height – among normally distributed variables. For erodible bank height, 19 of the remaining 20 variables passed the Pearson product-moment test with $p < 0.05$. When correlation coefficients of > 0.80 were taken to indicate a sufficient correlation with erodible terrace-bank height, 8 of the 19 independent variables were found to be sufficiently correlated with erodible terrace-bank height (Table 2). The stream network variables of the $\log_{10}$ transform of total stream-segment lengths and the non-transformed, maximum, stream-segment length showed the strongest correlations, with $r = 0.84$. The watershed morphometric variables of the natural log of drainage-basin perimeter and the $\log_{10}$ of drainage-basin length correlated with erodible terrace-bank height, with r values of 0.84 and 0.83, respectively. It is important to note that none of the land-cover or stream-reach morphology/sedimentology variables were sufficiently correlated with either floodplain width or erodible terrace-bank height.

For the dependent variable, $\log_2$ of the floodplain width, 11 of the 19 normally distributed independent variables passed the Pearson correlation test with a confidence interval of <0.05. Of these 11 remaining variables, all correlation coefficients (r-values) fell below 0.63. Therefore, correlates with r-values between 0.60 and 0.63 were arbitrarily selected as sufficient predictors of floodplain width, representing the best predictors for the dependent variable $\log_2$ of floodplain width. For this set, 9 of the remaining 11 independent variables sufficiently correlated with the $\log_2$ transform of floodplain width (Table 2). The stream network variable of the non-transformed maximum stream segment length showed the strongest correlation, with $r = 0.63$.

Modeled results

Bivariate regression models were created, first using different independent variables that were normally distributed and showed good correlation with the two dependent variables of floodplain width and erodible terrace-bank height. Of the independent variables, the $\log_{10}$ of total stream length was used to predict both erodible terrace-bank

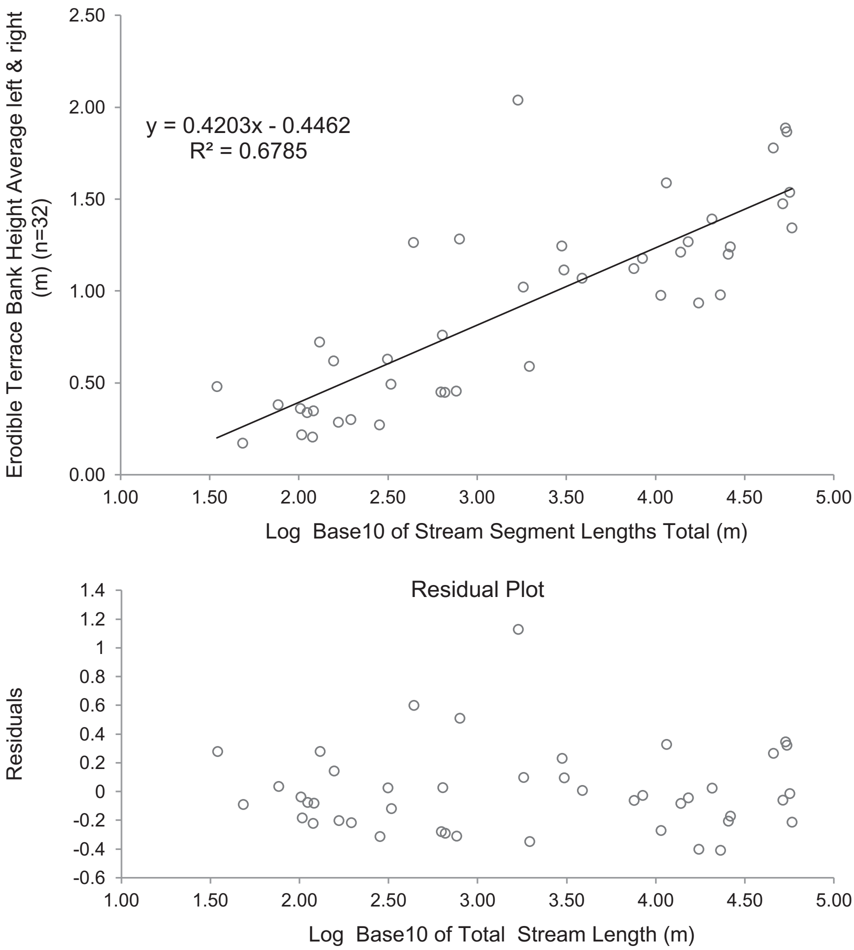

Figure 4. Scatter plot about the regressed prediction of erodible bank height predicted from the $\log_{10}$ of total stream length, and associated residuals.

height and floodplain width ($\log_2$ transformed). The $\log_{10}$ of total stream length was chosen because of the ease of replication using a GIS, and because total stream length represents a linear, scalar function that is most compatible with the prediction of the linear problem of bank erosion. Although the maximum, non-branching, stream-segment length was slightly better than total stream length as a predictor of floodplain width (r = 0.63 versus r = 0.61), total stream length was preferred as a logically more functional variable having greater likelihood of reproducibility outside of the study area. Furthermore, total stream length was more normally distributed than maximum stream segment length. The $\log_{10}$ of total stream length predicted terrace-bank heights using the following equation:

$$y = -0.4462 + (0.4203 \times x) \tag{1}$$

Here, y, is the modeled erodible terrace-bank height (untransformed, in meters) and x is the $\log_{10}$ of total stream length (m measured upstream from the midpoint of the particular modeled stream segment using a GIS. The r-squared value of this regression

Table 3. Total stream length modeled floodplain widths and erodible terrace bank heights for validation watersheds (results modeled by using total stream length as the independent variable).

		Widths (mm)	Heights (mm)
Deer lake drainage area 2.38 km²	Maximum	909.2	1205.4
	Minimum	166.1	145
	Mean	460	693.8
Thompson lake drainage area 3.83 km²	Maximum	976	1249.6
	Minimum	126.3	−25.9
	Mean	387.3	574.5
Bee tree drainage area 14.14 km²	Maximum	1434.6	1489.9
	Minimum	116.8	−75
	Mean	500.5	704.1
Cataloochee drainage area 127.43 km²	Maximum	2735.4	1892.5
	Minimum	116.8	−75.1
	Mean	551.6	712.5
Nantahala drainage area 134.42 km²	Maximum	2767.7	1899.8
	Minimum	116.8	−75.1
	Mean	421.6	622.3

model is 0.68, the F-statistic is 88.63, and $p < 0.0001$ (Figure 4). Using this model, the maximum erodible terrace-bank heights from all five validation watersheds ranged from 1.2 to 1.9 m, and their erodible bank-terrace heights ranged from 0.57 to 0.71 m (Table 3).

We also used the $\log_{10}$ of total stream lengths to model floodplain widths. The $\log_{10}$ of the total stream length predicted the $\log_2$ of floodplain widths as:

$$y = -3.9568 + (0.9720 \times x) \quad (2)$$

where y, is the $\log_2$ of floodplain width (m) and x is the $\log_{10}$ of total stream length (m), measured upstream from the midpoint of the modeled segment using a GIS. The r-squared value is 0.37, the F-statistic is 24.27, and $p < 0.0001$ (Figure 5). The antilog of the regression solution was subsequently factored into the GIS model for mathematical computation of total sediment yield.

Modeled maximum lateral migration rates of the two largest validation watersheds were 32.0 mm yr^{-1}, and the modeled maximum lateral migration rate of the smallest of the five validation drainage basins was 11.0 mm yr^{-1}. Mean migration rates of the five basins ranged from 5.0 to 6.0 mm yr^{-1}. All of these rates were calculated at the median date (91 years) of floodplain initiation presented in Table 4.

Sediment yield was modeled for these five validation drainage basins (Table 5) by multiplying the modeled lateral migration rate by the modeled erodible terrace-bank height by the length of each stream segment, and then summing those segment values for the entire drainage network. For the smallest drainage basin (Deer Lake) modeled, specific sediment yield was16.9 t km^{-2} yr^{-1}. Modeled specific yields for the largest basins, Cataloochee and Nantahala, were 24.5 and 25.0 t km^{-2} yr^{-1}, respectively. Again, these amounts represent the median erodible terrace-bank sediment yields midway between the range of floodplain initiation dates.

A multivariate approach using forward stepwise regression produced a better regression model of the floodplain width. However, for the erodible terrace-bank height, none of the other normally distributed independent variables improved the bivariate model presented above. Total stream length was forced into the model for previously mentioned reasons, and the best multivariate equation for the $\log_2$ of floodplain widths is:

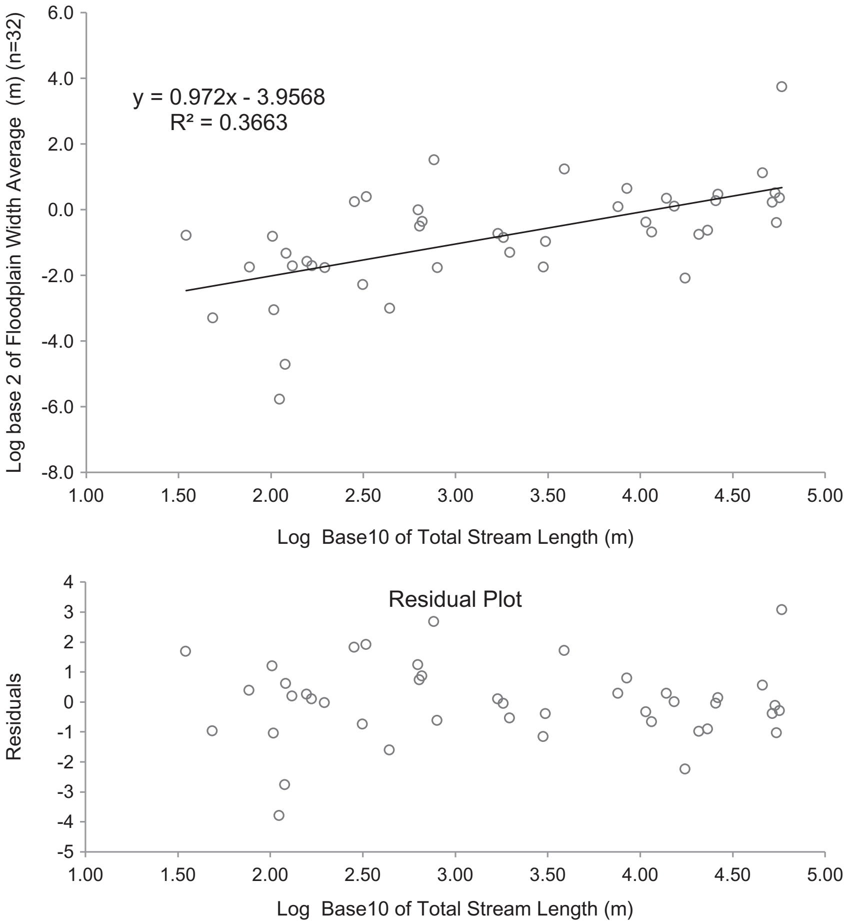

Figure 5. Scatter plot about the regressed prediction of $\log_2$ transformed floodplain widths predicted from the $\log_{10}$ of total stream length, and associated residuals.

$$y = -4.0823 + (0.5165 \times x_1) - (2.3085 \times x_2) \tag{3}$$

where y is the $\log_2$ of floodplain width (m), x_1 is $\log_{10}$ of total stream length (m) from the midpoint of the modeled segment, and x_2 is the modeled segment's basin-relief ratio (dimensionless basin total relief/basin length). The R^2 value for the multivariate regression model is 0.43, an improvement of 0.06 (6%) relative to the bivariate model that used only $\log_{10}$ of total stream length. The F-statistic of the multivariate model is not as strong as that of the $\log_{10}$ of total stream length bivariate model, at 15.23, but is highly significant ($p < 0.0001$).

Discussion

This study demonstrates that sediment yield solely from stream banks can be modeled at the watershed scale, based on empirical observations of floodplain widths. Statistical significance of our linear regression models and the similarity of their results to empirical observations from five independent studies in the region indicate that

Table 4. Lateral migration rates (mm/yr) of validation watersheds modeled with total stream length as the independent variable.

		Rate/112 years (mm/yr)	Rate/91 years (mm/yr)	Rate/70 years (mm/yr)
Deer lake drainage area	Maximum	8.1	10.6	13
2.38 km²	Minimum	1.5	1.9	2.4
	Mean	4.1	5.3	6.6
Thompson lake drainage	Maximum	8.7	11.3	13.9
area 3.83 km²	Minimum	1.1	1.5	1.8
	Mean	3.5	4.5	5.5
Bee tree drainage area	Maximum	12.8	16.7	20.5
14.14 km²	Minimum	1	1.4	1.7
	Mean	4.5	5.8	7.1
Cataloochee drainage area	Maximum	24.4	31.7	39.1
127.43 km²	Minimum	1	1.4	1.7
	Mean	4.9	6.4	7.9
Nantahala drainage area	Maximum	24.7	32.1	39.5
134.42 km²	Minimum	1	1.4	1.7
	Mean	3.8	4.9	6

Table 5. Total stream length modeled sediment yields (results modeled using $\log_{10}$ of total stream length as the independent variable).

		Total mass	Mass/112 Years	Mass/91 Years	Mass/70 Years
Deer lake drainage area	Tonnes	3660.05	32.68	40.22	52.29
2.38 km²	Tonnes/ km²/Year	1537.83	13.73	16.9	21.97
Thompson lake drainage	Tonnes	5692.47	50.83	62.55	81.32
area 3.83 km²	Tonnes/ km²/Year	1486.28	13.27	16.33	21.23
Bee Tree drainage area	Tonnes	23422.06	209.13	257.39	334.6
14.14 km²	Tonnes/ km²/Year	1656.44	14.79	18.2	23.66
Cataloochee drainage area	Tonnes	284151.88	2537.07	3122.55	4059.31
127.43 km²	Tonnes/ km²/Year	2229.87	19.91	24.5	31.86
Nantahala drainage area	Tonnes	306019.29	2732.32	3362.85	4371.7
134.42 km²	Tonnes/ km²/Year	2276.59	20.33	25.02	32.52

long-term sediment yield from stream-bank erosion can be effectively modeled from floodplain width proxies of bank erosion rates. Our model is limited to relatively small tributaries (< 20 km^2) of the Southern Blue Ridge, but these comprise a very large portion of the total drainage network in that region.

The lateral migration rates predicted from total stream length indicated that our model was producing realistic results (Table 4), based on comparisons to direct observations of bank erosion rates in the region by Harden et al. (2009) and Rhoades, O'Neal, & Pizzuto (2009). For the smallest validation watershed (Deer Lake at an area of 2.38 km²), the maximum rates of lateral migration range from 8.1 to 13.0 mm yr^{-1}.

The median modeled rate (based on median floodplain initiation 91 years BP) was 5.2 mm $yr^{-1}$for the three smaller watersheds and 5.6 mm yr^{-1} for the two larger watersheds. Considering all five validation watersheds, the mean modeled rates were 5.4 mm yr^{-1}, which compares favorably with the measurements of Harden et al. (2009) of 5.0 to 10.0 mm yr^{-1}, derived from bank-pin measurements in watersheds smaller than 980 km². The average of the maximum modeled rates of all five watersheds (depending on a date of floodplain initiation of 112 to 70 years before 2006) ranged from 15.8 to 25.2 mm yr^{-1}. This amount ranges from 9.9 to 15.8 mm yr^{-1} for the smallest three watersheds and 24.6 to 39.3 mm t km^{-2} yr^{-1} for the two largest. The maximum observed bank erosion rates of Harden et al. (2009) averaged 92.0 mm yr^{-1}, with mean rates of 20.0 mm yr^{-1}. Based on comparing channel migration based on sequential aerial imagery, Rhoades et al. (2009) found somewhat higher rates of 10.0 to 360.0 mm yr^{-1}, with an average of 40.0 mm yr^{-1}, but her watershed areas were not reported.

Modeled specific sediment yields (t km^{-2} yr^{-1}) also compared favorably with empirical observations from our validation sites (Table 6). For the smallest of the validation watersheds, Deer Lake (2.38 km²), modeled total sediment yields from banks were 16.90 t km^{-2} yr^{-1}, while the suspended sediment, observed by Royall (2003), was 29.7 t km^{-2} yr^{-1}. Though these amounts are somewhat similar, Royall indicated that his observed amounts were higher than expected based on what is typically observed in this region. Royall explained that evidence of mass wasting events along with the gradual destruction of sediment trapping debris dams (relics from the time of forest harvest), may have increased sediment yields through time for the Deer Lake basin. Royall recorded considerably lower sediment yields in two smaller watersheds. Thompson Lake (3.83 km²) and Bee Tree (14.14 km²) watersheds had measured sediment

Table 6. Comparative sediment yields between modeled and observed rates.

Bivariate models						
Author	Validation site	Drainage area (km²)	Observed mean sediment yield (t/km²/yr)	Total stream length estimated sediment yield (t/km²/yr)	Error from time of floodplain initiation range (t/km²/yr)	Note
Royall (2003)	Deer Lake, NC	2.38	29.7	16.9	±4.12	Total sediment yield
Royall (2000)	Thompson Lake, VA	3.83	8.3	16.33	±3.98	Total sediment yield
Simmons (1993)	Bee Tree Creek, NC	14.14	10.91	18.2	±4.44	Suspended sediment yield only
	Cataloochee River, NC	127.43	20.25	24.5	±5.97	Suspended sediment yield only
	Nantahala River, NC	134.42	14.95	25.02	±6.10	Suspended sediment yield only

yields of 8.3 and 10.9 t km^{-2} yr^{-1}, respectively, and our modeled sediment yields for these respective watersheds are marginally higher at 16.33 and 18.20 t km^{-2} yr^{-1}. For the next largest drainage basin of the Cataloochee Creek (127.43 km^2), the measured suspended sediment yield values were 20.25 t km^{-2} yr^{-1}, whereas the modeled yield results were slightly higher, at 24.50 t km^{-2} yr^{-1}. For the largest drainage basin of the five validation basins, Nantahala River (134.24 km^2), the observed suspended sediment yields were 14.95 t km^{-2} yr^{-1}and the modeled yield results were 25.02 t km^{-2} yr^{-1}. For the largest three watersheds, Bee Tree, Cataloochee, and Nantahala, the measured yields represent only suspended sediment load, so total sediment yield would be perhaps 10 to 25% higher, which is more consistent with our models. In contrast, the modeled estimates include both bank material, that inherently becomes suspended, and bed material, once it has been eroded.

Although the maximum watershed size of our observations was at a watershed area of about 20 km^2, the largest validation watersheds (up to 134 km^2) indicate that our models may be reliable somewhat beyond the limits of our observed data. However, at some watershed size, the linear relationship between total stream length and floodplain width/erodible bank height must break down because Leigh (2010) has observed that the main stem of the Little Tennessee River reflects different morphological behavior than its tributaries. That is, the main stem does not contain much of an historical terrace, so lateral migration rates are retarded in comparison to the tributaries. It is likely that there is a downstream time lag throughout the stream network in the progression of erosion and deposition of sediment, similar to what Trimble proposed as the “distributed sediment budget” (Trimble, 1993, p. 285). Thus, our predictions of erodible bank height and floodplain width probably break down somewhere in the range of 50 to 100 km^2 in basin area.

All 18 land-cover variables failed to make it into the model, which indicates that, in these catchments, basin land-cover is not an important factor influencing lateral erosion of the channel. Perhaps riparian vegetation cover is more influential, but we lack detailed riparian land-cover data for these basins. All but 1 of the 18 variables either failed the Spearman rank-order test or were drawn from a population with a non-normal distribution. Percent of deciduous forest was drawn from a population with a normal distribution, but failed the Pearson correlation test. Surprisingly, for these catchments, some bank and near-stream elements also failed incorporation into the models. These variables included dominant riparian land cover, bank slope, bank soil texture, percent bank vegetation cover, and banks that show active erosion. Another unexpected result was that variables that encapsulate size and slope at the basin scale, such as ruggedness number, drainage density, as well as many slope and gradient metrics, did not predict the dependent variables well. A full accounting of these variables and their correlative performance is included in Rogers (2011). This lack of ability to explain additional variability in the floodplain width perhaps implies that stochastic processes are operating, including feedback mechanisms and complex responses that are impossible to quantify, at least for this region. The fact that total stream length is the best correlate of floodplain width is logical, because it is a proxy for dominant discharge, and bank-erosion rates are expected to increase in the downstream direction as dominant discharge increases (Hooke, 1980). Lawler et al. (1999) found that bank-erosion rates were minimal in upper reaches in England’s far larger Swale-Ouse drainage systems because of lower bank heights relative to those of middle and lower reaches. As with our findings, an exception may pertain to stream networks that migrate into historic terraces.

Walter and Merits (2008) and Pizzuto and O'Neal (2009) found that the breaching of former mill dams have had a critical influence on channel form in the eastern USA. However, Pizzuto and O'Neal indicated that increased rates of bank erosion were not completely explained by mill dam removal and suggested that differences in lateral erosion rates may be related to local geomorphic processes, stochastic variations in bank-erosion rates, mill-dam effects not assessed by simple backwater computations, and/or changes in land cover. Certainly, the stochastic nature of both natural and human processes within the basin makes predicting bank-erosion rates difficult. For instance, land-cover change through the period of this study has been temporally dynamic in the study area (Kirk, 2009). In a mountainous area such as the study area, complex response surely plays a role in sediment yields, as well as rates of lateral migration.

Potential errors not captured by our modeled floodplain widths and erodible terrace-bank heights are associated with channel form, stream processes, and errors associated with the validation studies. Variation in channel form may explain some of the variation in our models. This includes channel widening or narrowing through time, thereby skewing lateral migration rates. Leigh (2010) found a statistically significant relationship between wider channels and forested reaches, as well as narrower channels and pastured reaches. Other channel characteristics not captured by the model are human influences such as bank riprapping and channelization. Stream processes captured by the model include unidirectional but not bidirectioinal stream migration. The past directionality of stream migration can be difficult to determine, especially in relatively straight reaches. However, in many of our observations in distinct meander bends, the migration appears to have been unidirectional. Thus, this modeled yield represents a net migration into the portion of the bank above the floodplain.

Although we sought independent estimates of sediment yield to test our bank sediment yield model, we must acknowledge that there may be error within the validation studies. Analytical errors involved with the studies of Royall (2000, 2003) and Simmons (1993) are not reported, but the authors did mention concerns of error. Simmons mentioned a limitation with regard to the time of the study, which was characterized by higher than normal stream discharge (1970–1979 study versus 30-year averages from1950 to 1979), but it is uncertain how this would have affected bank-erosion rates in the fully forested watershed he studied. Royall (2000) indicated that the primary factor controlling sediment yield fluctuation was related to long-term hydrologic discharge from variation in precipitation. However, during the 29 years (1965–1995) of lake sediment accumulation, when these discharges fluctuated broadly, mean discharges remained moderate. Royall (2003) addressed another potential source of error in assessing 50 years of lake sediment accumulation: at the end of the life of the dam, dam failure might result from a heavy rainfall event. He suggested that this meteorological event, as well as the destruction of coarse woody debris dams in conjunction with lakeside soil-disturbing construction activities, may have led to the unusually high sediment yields observed in his study, and noted that, methodologically, such stochastic processes are more obscure when relating sediment yields.

Despite the potential errors discussed above, it is apparent that the use of floodplain width as a proxy for bank-erosion rates is a viable approach that produces reasonable results that are remarkably close to values derived from very different empirical methods. A clear advantage of using the floodplain width as a proxy is that it does not necessitate long-term monitoring to develop a sound database. Measurement of floodplain widths throughout the drainage network also provides a spatially extensive database that is difficult to achieve with monitoring efforts using erosion pins and bedload and suspended load samples. The only limitation to our method is the establishment of an accurate age of floodplain formation, but such age estimates already have been established in many regions (Jacobson & Coleman, 1986; Knox, 1977, 1987) and, with refined methods of radiocarbon and luminescence dating, it has become relatively easy to establish reliable alluvial chronologies.

Conclusion

Excess sediment is one of the primary stream pollutants in the world today. Understanding sources and dynamics of sediment is critical to minimize its negative impacts. This study has isolated a relatively small geographic region to attempt to answer, in part, a critical question. That is, how much sediment do stream banks add to sediment yields? We addressed this question using floodplain width as a proxy for long-term lateral erosion rates and using observations of bank heights incorporated in a GIS model that predicts sediment yields from banks within tributaries of the Southern Blue Ridge.

This study estimated bank-derived sediment yield based on comparisons with known sediment yields in these fully forested basins where the stream channel (mainly banks) is the primary source of sediment. Our modeled rates of lateral erosion fall within the ranges of the observed rates in the region, based simply on total stream length as the independent variable. The scale-dependent observations can be applied over an entire small watershed with reasonable results. However, there is a limit to the size of drainage basins in which our modeled results are reliable, given that the model relies on data from watersheds limited to 20 km^2. Addition of a second predictor, the drainage-basin relief ratio, in multiple regressions added about 6% to the estimate of lateral erosion rates, but required significantly more computation in the GIS model.

Our model provides an important step towards a better understanding of the contribution that bank erosion delivers to total sediment yield in watersheds of the Southern Blue Ridge. Indeed, our findings indicate that it is reasonable to expect most of the sediment derived from fully forested basins to be from bank erosion, as other sources of sediment are few under such land-cover conditions. In addition, our results suggest that observed sediment yields in excess of 10 to 25 t km^{-2} yr^{-1}in watersheds smaller than 100 km^2 almost certainly involve sources of sediment other than stream banks. Furthermore, our results indicate that bank erosion operates independently from basin land-cover conditions and perhaps even independently from riparian land-cover conditions.

Acknowledgments

This research was funded by the National Science Foundation (DEB-0218001 and DEB-0823293) and sponsored by the US Forest Service Southern Research Station in cooperation with the University of Georgia (SRS 04-CA-1133 0140-096). The University of Georgia Research Foundation, Inc. provided administrative support. Jim Vose, Brian Kloeppel, Carol Harper, and staff of the Coweeta Hydrologic Laboratory provided a great deal of support during data collection. Brad Suther, Jason Meadows, Katie Price, Leslie Martin, Karin Rogers, and Hugh Dana directly assisted in data collection. Rhett Jackson, Marguerite Madden, Steve Patch, and Rick Smith provided advice and assistance at key moments. Several institutions were critical in the completion of this research: The University of Georgia Graduate School, The U.S.D.A. Forest Service, Coweeta Long Term Ecological Research program, and The University of North Carolina Asheville. We would like to thank an anonymous reviewer as well as Drs Dan Royall, Tim Stott, and Allan James for their feedback, which improved the quality of this article. We acknowledge all of the private landowners who graciously allowed access to their property during the data collection phase of this study.

References

Ayers, H. B., & Ashe, W. W. (1904). *The Southern Appalachian forests* (U.S. geological survey professional paper No.37). Washington, DC: U.S Geological Survey.

Eller, R. D. (1982). *Miners, millhands, and mountaineers: Industrialization of the Appalachian South 1880–1930*. Knoxville, TN: University of Tennessee Press.

Fitzpatrick, F., Waite, I., D'Arconte, P., Meador, M., Maupin, M., & Gurtz, M. (1904). *Revised methods for characterizing stream habitat in the national water-quality assessment program* (Report No. 98-4052). Raleigh, NC: U. S. Geological Survey, Water-Resources Investigations.

Glenn, L. C. (1911). *Denudation and erosion in the Southern Appalachian region and the Monongahela Basin* (U.S. geological survey professional paper no. 72). Washington, DC: Government Printing Office.

Harden, C. P. (2004). Fluvial response to land-use change in the Southern Appalachian region: A century of investigation. *Physical Geography, 25*, 398–417.

Harden, C., Foster, W., Morris, C., Chartrand, K., & Henry, E. (2009). Rates and processes of streambank erosion in tributaries of the Little River, Tennessee. *Physical Geography, 30*, 1–16.

Harrelson, C. C. Rawlins, C. L., & Potyondy, J. P. (1994). *Stream channel reference sites: An illustrated guide to field technique* (USDA general technical report RM-245). Fort Collins, CO: U.S. Forest Service Rocky Mountain Forest and Range Experiment Station.

Homer, C., Dewitz, J., Fry, J., Coan, M., Hossain, N., Larson, C., Herold, N., … Wickham, J. (2007). Completion of the 2001 national land cover database for the conterminous United States. *Photogrammetric Engineering and Remote Sensing, 73*(4). Retrieved October 15, 2005, from http://www.mrlc.gov/nlcd2001.php

Hooke, J. M. (1979). An analysis of the processes of river bank erosion. *Journal of Hydrology, 42*, 39–62.

Hooke, J. M. (1980). Magnitude and distribution of rates of river bank erosion. *Earth Surface Processes, 5*, 143–157.

Jacobson, R. B., & Coleman, J. C. (1986). Stratigraphy and recent evolution of Maryland piedmont flood plains. *American Journal of Science, 286*, 617–637.

Kaufmann, P. R., Levine, P., Peck, D. V., Robison, E. G., & Seeliger, C. (1999). *Quantifying physical habitat in wadeable streams* (US Environmental Protection Agency Report EPA 620/R-99/003). Washington, D.C.: Environmental Monitoring and Assessment Program, National Health and Environmental Effects Research Laboratory, Office of Research and Development.

Kelsey, H. M. (1982). Influence of magnitude, frequency, and persistence of various types of disturbance on geomorphic form and process. In F. J. Swanson, E. J. Janda, T. Dunne, & D. N. Swanson (Eds.), *Sediment budgets and routing in forested drainage basins* (pp.150–153). Washington, DC: USDA Forest Service General Technical Report PNW 141.

Kirk, R. W. (2009). *Land use and terrestrial carbon storage in Western North Carolina from 1850–2030: A historical reconstruction and simulation study* (Unpublished doctoral dissertation). University of Minnesota, Twin Cities, MN.

Knighton, D. (1998). *Fluvial forms and processes: A new perspective*. London: Arnold.

Knox, J. C. (1977). Human impacts on Wisconsin stream channels. *Annals of the Association of American Geographers, 67*, 323–342.

Knox, J. C. (1987). Historical valley floor sedimentation in the upper Mississippi Valley. *Annals of the Association of American Geographers, 77*, 224–244.

Lawler, D. M. (1993). The measurement of river bank erosion and lateral change: A review. *Earth Surface Processes and Landforms, 18*, 777–821.

Lawler, D., Grove, J., Couperthwaite, J., & Leeks, G. (1999). Downstream change in riverbank erosion rates in the Swale-Ouse system, northern England. *Hydrological Processes, 13*, 977–992.

Leigh, D. S. (2010). Morphology and channel evolution of small streams in the southern Blue Ridge Mountains of western North Carolina. *Southeastern Geographer, 50*, 397–421.

Leigh, D. S., & Rogers, J. C. (2007). *Geomorphic analysis of stream channels in the Southern Appalachian Highlands* (Technical report of U.S. forest service and university of Georgia, geomorphology laboratory technical research report no. 3). Athens, GA: University of Georgia.

Leigh, D. S., & Webb, P. A. (2006). Holocene erosion, sedimentation, and stratigraphy at Raven Fork, Southern Blue Ridge Mountains, USA. *Geomorphology, 78*, 161–177.
Madej, M. A., & Ozaki, V. (1996). Channel response to sediment wave propagation and movement, Redwood Creek, California, USA. *Earth Surface Processes and Landforms, 21*, 911–927.
Meade, R., Yuzyk, T. R., & Day, T. J. (1990). Movement and storage of sediment in rivers of the United States and Canada. In M. G. Wolman & H. C. Riggs (Eds.), *Surface water hydrology: The geology of North America*, vol. O–1 (pp. 255–280). Boulder, CO: Geological Society of America
Montgomery, D. R. (1999). Process domains and the river continuum. *Journal of the American Water Resources Association, 35*, 397–410.
Murgatroyd, A. L., & Ternan, J. L. (1983). The impact of afforestation on stream bank erosion and channel form. *Earth Surface Processes and Landforms, 8*, 357–370.
Nanson, G. C., & Hickin, E. J. (1986). A statistical analysis of bank erosion and channel migration In Western Canada. *Geological Society of America Bulletin, 97*, 497–504.
National Climate Data Center (2011). *Climatetography of the United States, No. 84, 1981–2010*. National Oceanic and Atmospheric Administration, Retrieved November 1, 2011, from ftp://ftp.ncdc.noaa.gov/pub/data/normals/1981-2010/products/station/USC00312102.normals.txt
Petras, I. (2003). *Basin1 ArcView [Script/extension]*. Environmental Systems Research InstituteRedlands, CA: Department of Water Affairs & Forestry, Pretoria, South Africa.
Pizzuto, J. E., & O'Neal, M. O. (2009). Increased mid-20th century river bank erosion rates related to the demise of mill dams, South River, Virginia. *Geology, 37*, 19–22.
Price, K., Jackson, C. R., & Parker, A. J. (2010). Variation of surficial soil hydraulic properties across land uses in the southern Blue Ridge Mountains, NC, USA. *Journal of Hydrology, 383*, 256–268.
Price, K., & Leigh, D. S. (2006a). Stream morphological and sedimentological response to human impact in the Southern Blue Ridge Mountains, USA. *Geomorphology, 78*, 142–160.
Price, K., & Leigh, D. S. (2006b). Comparative water quality of lightly and moderately impacted streams in the Southern Blue Ridge Mountains, USA. *Environmental Monitoring and Assessment, 120*, 269–300.
Reid, L. M. (1993). *Research and cumulative watershed effects* (U.S. department of agriculture general technical report PSW-GTR-141). Albany, CA: Pacific Southwest Research Station, Forest Service.
Reid, L. M., & Dunne, T. (1996). *Rapid evaluation of sediment budgets*. Reiskirchenberg: Catena Verlag GMBH.
Rhoades, E. L., O'Neal, M. A., & Pizzuto, J. E. (2009). Quantifying bank erosion on the South River from 1937 to 2005, and its importance in assessing Hg contamination. *Applied Geography, 29*, 125–134.
Robinson, G. R., Jr., Lesure, F. G., Marlowe, II. J. I., Foley, N. K., & Clark, S. H. (1992). *Bedrock geology and mineral resources of the Knoxville 1°x2° Quadrangle, Tennessee, North Carolina, and South Carolina (Digital Version)* [Data set]. Retrieved October 3, 2011, from http://pubs.usgs.gov/of/2004/1075/
Rogers, J. C. (2011). *Stream bank erosion of tributaries in the Southern Blue Ridge Mountain* (Unpublished master's thesis). University of Georgia, Athens, Georgia. Retrieved from https://getd.libs.uga.edu/pdfs/rogers_james_c_201112_ms.pdf
Roosevelt, T. (1902). *Message from the president of the United States transmitting a report from the secretary of agriculture in relation to the forests, rivers, and mountains of the southern Appalachian region*. Washington, D.C.: Government Printing Office.
Royall, D. (2000). Lake-sediment-based evaluation of recent sediment yield from a small Appalachian watershed, Thompson Lake, Virginia. *Physical Geography, 21*, 68–88.
Royall, D. (2003). A 50-year record of sediment yield and sources from Wolf Branch, North Carolina. *Professional Geographer, 55*, 356–371.
Shreve, R. L. (1967). Infinite topologically random channel networks. *Journal of Geology, 75*, 178–186.
Simmons, C. E. (1993). *Sediment characteristics of North Carolina streams, 1970–79* (United States geological survey, water supply paper 2364). Raleigh, NC: United States Geological Survey.

Simon, A., & Castro, J. (2003). Measurement and analysis of alluvial channel form. In G. M. Kondolf & H. Piégay (Eds.), *Tools in fluvial geomorphology* (pp. 291–322). West Sussex: John Wiley and Sons.

Simon, A., Curini, A., Darby, S., & Langendoen, E. (2000). Bank and near-bank processes in an incised channel. *Geomorphology, 35*, 193–217.

Swank, W. T., Vose, J. M., & Elliott, K. J. (2001). Long-term hydrologic and water quality responses following commercial clear cutting of mixed hardwoods on a southern Appalachian catchment. *Forest Ecology and Management, 143*, 163–178.

Thorne, C. R. (1982). Processes and mechanisms of riverbank erosion. In R. D. Hey, J. C. Bathurst, & C. R. Thorne (Eds.), *Gravel-bed rivers: Fluvial processes, engineering and management* (pp. 227–271). West Sussex: John Wiley and Sons.

Trimble, S. W. (1993). The distributed sediment budget model and watershed management in the Paleozoic Plateau of the upper Midwestern United States. *Physical Geography, 14*, 285–303.

Walling, D. E., & Fang, D. (2003). Recent trends in the suspended sediment loads of the world's rivers. *Global and Planetary Change, 39*, 111–126.

Walter, R. C., & Merritts, D. J. (2008). Natural streams and the legacy of water-powered mills. *Science, 319*, 299–304.

Wolman, M. G. (1959). Factors influencing erosion of a cohesive river bank. *American Journal of Science, 257*, 204–216.

Yarnell, S. L. (1998). *The Southern Appalachians: A history of the landscape* (General Technical Report, SRS-18). Asheville, NC: U.S. Department of Agricultural Forest Service Southern Research Station.