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Downstream geomorphic variation and local bedrock influence of a steep transitional river: Blue Ridge to Piedmont, South Carolina

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Spatial patterns in geomorphic variations are examined in a river transition from the Southern Blue Ridge to the Piedmont physiographic regions. Downstream hydraulic geometry (DHG), fining of bed material, and changes in reach-scale channel-bed morphology (bedforms) were sampled and analyzed. DHG power functions were well developed ($r^2 > 0.75$ for channel area, width, and depth). Bed material showed a general downstream exponential decrease in caliber (from 940 mm to sand). However, variations within the general downstream trends reflect local variations and a rapid transition in hydraulic variables and bedforms at a key zone with substantial tributary inputs, decreases in slope, and a punctuated longitudinal profile. Structurally controlled bedrock knickpoints are associated with anomalous spatial patterns of bedforms and can be distinguished through relationships between dimensionless sediment transport capacity and sediment supply using a sediment regime diagram that is independent of drainage area. Plots of relative grain submergence (R/D_{84}), relative form submergence (R/H), Darcy–Weisbach friction factor (f), and slope vs. area also reveal trends that suggest local factors that are missed by downstream hydraulic progressions. The findings of this study corroborate the utility of scale-independent methods, especially in mountain or transitional environments where fluvial controls may be longitudinally sporadic.

Keywords: Blue Ridge; bedforms; mountain rivers; step-pool; pool-riffle; forced morphology; sediment regime diagram; downstream hydraulic geometry; downstream fining

Introduction

Many well-established principles of fluvial geomorphology, including downstream hydraulic geometry (DHG), were developed with relatively low-gradient rivers. The conceptual theory of DHG suggests that systematic downstream changes in channel-top width (w), mean flow depth (d), and mean velocity (v) can be expressed as simple power relationships with discharge (Leopold & Maddock, 1953). DHG relationships assume that alluvial channels adjust to changes in discharge or sediment supply toward an approximate equilibrium state (Wohl, Kuzma, & Brown, 2004). This general relationship has held up well in regional-scale studies (Faustini, Kaufmann, & Herlihy, 2009) and is widely used in research and river management. However, studies on DHG of steep mountain rivers have produced mixed results, with conclusions of both well-developed and nonexistent DHG relationships (Wohl, 2004). Challenges to DHG have

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been made based on highly variable results derived from hyper resolution studies (Carbonneau, Fonstad, Marcus, & Dugdale, 2012). Similarly, results of general downstream fining trends in bed material size are also highly variable in mountain environments (Gomez, Rosser, Peacock, Hicks, & Palmer, 2001).

Recent research has focused on methods suited for characterizing steep rivers, including the use of sediment and hydraulic data to establish standard geomorphic principles in such environments (Montgomery & Buffington, 1997; Wohl & Merritt, 2008). Montgomery and Buffington (1997) proposed a classification for steep river channel-bed morphology (bedforms) as a function of the relationship between transport capacity (e.g., total shear stress) and sediment supply. Although their classification was framed in the context of an idealized downstream progression, the concept is used as the foundation for methods independent of drainage area. Thompson, Croke, Ogden, and Wallbrink (2006) and Wohl and Merritt (2008) further supported the idea that mountain river morphologies can be distinguished with different analyses of hydraulic and sedimentological variables, independent of drainage area.

These methods can be used to evaluate local variability in channel morphology in steep or transitional river environments, where general downstream trends fail to do so. The body of literature specific to steep rivers and geomorphic transition zones (e.g., from the mountains into the piedmont) reveals variable results due to the localized nature of geomorphic influences in such environments (Fryirs & Brierley, 2010; Fryirs, Brierley, Preston, & Kasai, 2007). The research here utilizes methods that can be applied to environments with complexities in DHG, bed material size, and bed material arrangement (Montgomery & Buffington, 1997; Thompson et al., 2006; Wohl et al., 2004; Wohl & Merritt, 2008). Mountain rivers and transition zones represent a diversity of aquatic ecosystem habitat types within a relatively small range of drainage areas (Church, 2002; Price & Leigh, 2006). Further, channel habitat associated with local variability may exhibit varying degrees of response to disturbance (Montgomery & Buffington, 1998). Thus, understanding the hydraulic, sedimentological, and landscape-scale context of geomorphic variation is desirable for river management.

Further, research in fluvial geomorphology from the Southern Appalachians is lacking compared to other mountainous regions (Harden, 2004), especially with regard to channel form and processes. The literature that exists (i.e., Leigh, 2010; Leigh & Webb, 2006; Price & Leigh, 2006) is often from research in basins that drain to the Tennessee River, which have different basin characteristics, especially lower gradients, than those that drain the southern edge of the Blue Ridge escarpment toward the Atlantic Ocean (Haselton, 1974).

The goal of this research was to explore geomorphic variations through the course of a transitional river in the Blue Ridge physiographic region of South Carolina. First, general downstream trends were explored using models of DHG, downstream fining of bed material, and the downstream progression of bedforms. Second, methods independent of drainage area were used to evaluate variability in the general downstream trends. These methods quantify the key influences of geomorphic variation in the transition zone with hydraulic and sedimentological data. Finally, the geomorphic influence of bedrock knickpoints – present throughout the transition – was explored at a reach and segment scale.

A brief review of the literature in three distinct categories – DHG, downstream fining of bed material, and reach-scale channel bedforms – as it applies to mountain and transitional fluvial environments is given as the foundational context for this research.

Downstream hydraulic geometry

DHG assumes that there is sufficient power in bankfull flows for channel morphology to adjust to systematic changes in discharge downstream (Leopold & Maddock, 1953). This assumption is most valid in alluvial rivers, but streams geomorphically forced by external influences – e.g., woody debris, colluvial inputs, valley constrictions, bedrock knickpoints, human distrubance – can exhibit high variability in channel geometry (Clark & Wilcock, 2000; Wohl, 2004; Wohl et al., 2004). These geomorphic influences are common in mountain environments. Wohl (2004) hypothesized a threshold of excess stream power (Ω/D_{84} = 10,000 kg s^{-3}) at which well-developed DHG relationships commence in mountain rivers. This helps to address the assumption of a deformable boundary adjusting to regular flows. However, the assumptions of DHG are also dependent on scale and resolution of the study (Carbonneau et al., 2012).

Downstream fining

Fining of bed material with distance downstream is related to hydraulic variables, particularly slope (Knighton, 1998), abrasion of materials at the channel bed (Kodama, 1994a, 1994b), and hydraulic sorting and transport of bed material (Ferguson, Hoey, Wathen, & Werritty, 1996; Wilcock & McArdell, 1993). Others have explored variations within the trend of downstream fining. Best (1988), Rice (1998), and Rădoane, Rădoane, Dumitriu, and Miclăuş (2008) have revealed the importance of tributary inputs to downstream textural variations in bed material, while other studies have revealed abrupt fining trends associated with local controls of slope (Ferguson & Ashworth, 1991; Surian, 2002). Certain homogeneous conditions within a basin may develop a continuous downstream fining trend and a rapid gravel-to-sand transition (Gomez et al., 2001). Other studies report that bimodal sediment distributions and selective transport can generate a gradual gravel–sand transition (Rădoane et al., 2008; Singer, 2008). Further, systematic downstream coarsening of bed material is observed in headwater channels in Washington state until a threshold of drainage area at which downstream fining commences (Brummer & Montgomery, 2003; Leigh, 2010). Bed material dynamics can be highly variable in mountain environments; thus, analyzing bed material trends in this study is important for assessing geomorphic features through the transition zone.

Bedforms

The hydraulic forces of moving water and erodible sediment on the channel bed create morphological bedforms in the channel at the reach scale (Knighton, 1998). Montgomery and Buffington (1997) developed a classification for mountain streams in which they identified progressive changes in bedform types (Figure 1). Their classification proposed that morphologies of mountain channels with movable beds are a function of the relationship between transport capacity (e.g., total shear stress), which typically decreases downstream, and sediment supply, which generally increases downstream.

Bedform types in the Montgomery and Buffington (1997) classification typically follow a downstream progression: cascades with very coarse bed material and little organization; step-pools with coarse bed material organized into relatively evenly spaced steps and plunge pools; plane-beds characterized by little channel bed topography and uniform bed material size; pool-riffles with fine bed material arranged into a

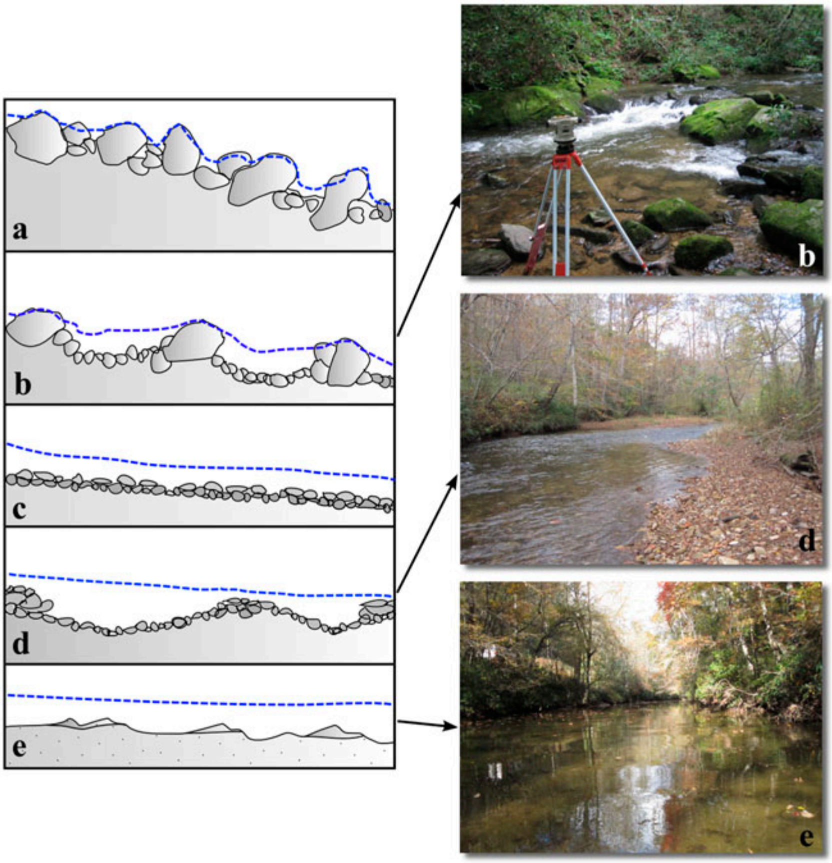

Figure 1. Simplified bedform types. In order of typical downstream progression: (a) cascade, (b) step-pool, (c) plane-bed, (d) pool-riffle, and (e) sand bed, and expressed as an *infilled* morphology in this paper. Photographs are from the Middle Saluda River. Adapted from Montgomery and Buffington (1997).

series of riffles followed by pools; and a sand bed sometimes arranged into dunes or ripples. The classification is based on a progression of hydraulic and sediment variables that tend to change systematically downstream, even though channel types may not necessarily progress downstream in a particular river due to local conditions, including gradient discontinuities, sediment and tributary inputs, sediment storage, large woody debris, and human impacts (Montgomery & Buffington, 1997; Thompson et al., 2006; Wohl et al., 2004; Wohl & Merritt, 2008).

Wohl and Merritt (2008) analyzed data from mountain river reaches around the world, giving value ranges to critical sediment and hydraulic variables associated with Montgomery and Buffington (1997) bedform types. Thompson et al. (2006) found intermediate morphologies in their statistical analysis that reflect slight variations in process, form, and lithologies and fall somewhere between the bedform types proposed by Montgomery and Buffington (1997). In some cases, bedforms are influenced by variables independent of a downstream hydraulic progression, such as large woody debris, bedrock knickpoints, or abrupt changes in gradient. Montgomery and Buffington

(1997, 1998) described this as a *forced* morphology; *forced* can be applied to any morphological type (e.g., forced step-pool). Thompson et al. (2006) described an *in-filled* morphology, which exhibits a featureless sand bed that appears forced by an external condition, such as increased sediment supply or bedrock control of slope. In this paper, *forced* is used to describe any morphology influenced by a force other than the downstream progression of hydraulics, while *infilled* will be used to describe the specific type of forced morphology with a featureless sand bed that appears to overwhelm the channel bed. Figure 1e is an example of a forced and infilled morphology. Forced morphologies are important to recognize because they imply anomalous forms not predicted by downstream models, which has implications for response to disturbance (Montgomery & Buffington, 1998; Fryirs et al., 2007).

Physical setting

The Middle Saluda River emerges from the Blue Ridge Escarpment in northern South Carolina and flows through a transition zone between the Blue Ridge and Piedmont physiographic regions. The Middle Saluda River is located in Greenville County, South Carolina, United States of America. The river heads in the Blue Ridge Escarpment and flows for 31 km through the study watershed. The drainage area of the study watershed is 110 km^2. Folded gneiss, augen gneiss, and schist of different formations dominate the watershed (Garihan, 2005). A series of faults extend in a general WSW to ENE direction. The trellised drainage pattern is structurally controlled by the faults, joints, and foliation trends, with tributaries approaching the main stem perpendicularly at confluences (Haselton, 1974). The river descends steeply (average gradient 0.06) from the head through a confined valley, followed by lower gradient valleys with alternating floodplain pockets. At some locations, the main stem flows through narrow gaps across the structural ridges where channels are laterally confined (Figure 2). The lower section of the river has a mean gradient of 0.003, which is substantially less than that of the upstream section but still considered "steep" (Montgomery & Buffington, 1997). The transition from very steep gradients to lower gradients is described here as the "transition zone" and is a focus of this study. The greatest elevation in the watershed is 1152 m (amsl). Elevation of the river bed ranges from approximately 888 m on the escarpment to 300 m at the outlet.

Average annual precipitation in the basin is higher than in most regions of the southeastern United States of America, ranging from 192 cm on the escarpment to 151 cm near the outlet (SC State Climate Office, 2012). Snowfall is mostly concentrated in the upper watershed at the highest elevations. Monthly precipitation is relatively uniform throughout the year. Land cover in the watershed is 92% forest, 6% agricultural or recently deforested, 1% urban, and 1% water and rock outcrop features (Southeast Gap Analysis Project, 2011). Impoundments in the watershed are minimal and are mostly located near the headwaters of tributaries. The USGS streamflow gage, Middle Saluda River at Cleveland (02,162,350), is located at a drainage area of 52 km^2. Mean annual flows at the gage range from 0.8 m^3 to 2.5 $m^3\ s^{-1}$ for the period of record.

Research on the Middle Saluda River watershed is important because of: (1) increased interest in management, restoration, and protection of rivers in the region; (2) improved knowledge of mountain river processes, which has generated a need to modernize the conceptual understanding of this system; and (3) the scarcity of literature on Southern Appalachian rivers relative to rivers in other mountainous regions. The study

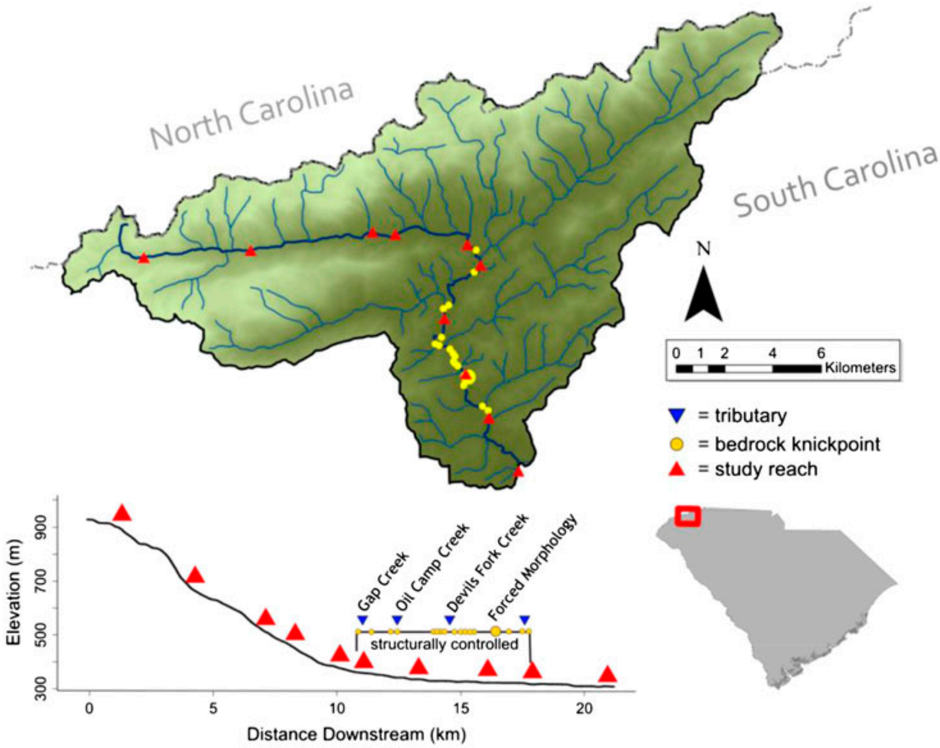

Figure 2. Study watershed overview of the Upper Middle Saluda River, South Carolina, including locations of study reaches, bedrock knickpoints, and large tributaries from a map view and along the longitudinal profile (distance downstream is from uppermost headwater channel).

reach of the Middle Saluda River is accessible by roads, trails, and canoe, offering an excellent opportunity to collect field data through the transition zone. Further, the watershed is currently affected by little human impact relative to the surrounding area.

Methodology

Field data collection

Studies by Montgomery and Buffington (1997), Thompson et al. (2006), and Wohl and Merritt (2008) served as a guide for channel geometry, bed material, and bedform field data collection in this study. Collection of field data was designed to sample channel bed material and survey channel morphology systematically downstream. Sites were chosen to represent a variation of known influences on channel morphology and sediment size, including drainage area, slope, valley types, location of tributaries, and proximity to bedrock knickpoints. The field measurements made and calculations used in the study are given in Table 1.

Ten study reaches were sampled after identifying river reaches ~ 10 – 20 times channel width with consistent morphology (Wohl & Merritt, 2008), except for two shorter reaches where vegetation limited surveying opportunities. Cross-section sites were chosen that characterized the reach and were surveyed with a rod and level.

Table 1. Explanation of variables and calculations.

Variable (units) – *symbol*	Explanation	Method
BASIN DATA		
Drainage Area (km^2) – *DA*	Upstream drainage from reach	GIS
Valley Width (m) – *VW*	Width of valley (i.e., floodplain) at reach	GIS
CROSS-SECTION		
Width (m) – *w*	Cross-section width at bankfull	Survey
Depth (m) – *d*	Cross-section average depth at bankfull	Survey
Area (m) – *area*	$area = w \times d$; area of cross section at bankfull	Calculation
Entrenchment ratio – *ER*	$\frac{W_{2dmax}}{w}$; W_{2dmax} is width at 2 × max. bankfull depth	Calculation
Wetted perimeter – *P*	at bankfull	Calculation
Hydraulic radius – *R*	*area*/*P* at bankfull	Calculation
HYDRAULIC VARIABLES		
Gradient ($m\ m^{-1}$) – *S*	Slope of energy grade line at bankfull – bankfull indicators	Survey
Velocity; manning's equation ($m\ s^{-1}$) – *v*	$\frac{R^{2/3}S^{1/2}}{n}$ *n* is Manning's roughness coefficient (estimated)	Calculation
Discharge ($m^3\ s^{-1}$) – Q_{bkf}	$area \times v$; at bankfull	Calculation
Cross sectional stream power ($kgm\ s^{-3}$) – Ω	ρgQS; ρg is density of fluid × acceleration due to gravity, constant 9800; at bankfull	Calculation
Mean boundary shear stress (pascals) – τ	ρgRS; at bankfull	Calculation
Darcy-Weisbach friction factor (dimensionless) – *f*	$\frac{8gRS}{v^2}$ at bankfull, estimator of roughness.	Calculation
BED MATERIAL		
Representative particle size (mm) – D_x	Size at which x percent of particles are smaller on cumulative frequency distribution	Calculation
Sand depth (m) – *sand*	Depth of sand in pools	Probe
BEDFORMS		
Amplitude (m) – *H*	i.e., crest of step to bottom of pool, vertical measurement	Survey
Wavelength (m) – *L*	Pool to pool, or crest to crest. Longitudinal measurement	Survey

Bankfull markers were identified in the field using indicators discussed in the United States Forest Service tutorial (e.g., slope breaks, cobble lines, and undercut banks; USFS, 2008). Prior analysis of USGS gage data (Feaster, Gotvald, & Weaver, 2009) was used to constrain the magnitudes of estimates of bankfull variables. Bankfull markers were surveyed longitudinally to attain bankfull water surface slope, which was used to approximate slope of the energy grade line for use in slope-area computations of bankfull discharge. Longitudinal measurements were also taken along the thalweg in the channel, from which bedform amplitude and wavelength parameters were calculated. Bed material was sampled using a surface grid sampling method (Wolman, 1954). The coarsest active bed material in each reach was sampled at all 10 cross sections, plus three additional sites to better characterize bed material. The sand component of the channel bed (fine mode) was quantified by probing the depth of sand in pools, which visual observations suggested increase downstream.

Data analysis

Cross sections were analyzed using a third-party spreadsheet program (NRCS, 2012a). User inputs include cross-sectional stations, elevations, roughness (Manning's *n*), and reach slope. Output includes a suite of hydraulic variables associated with a range of stage values within the cross section, including *area*, P, R, w, d, v, τ, f, and Q_{bfk}, as defined in Table 1. The cross section was selected to match bankfull stage observed in the field to provide estimates of hydraulic variables associated with bankfull flows.

Careful attention was given to estimating bankfull conditions using the slope-area method, which requires estimates of hydraulic roughness. Manning's roughness (n) was given particular attention so that estimates of hydraulic calculations were as accurate as possible. Several methods were used in estimating roughness. Barnes (1967) provided a visual-comparison method of estimation based on photographs with measured values of roughness. Chow (1959) used an iterative method, adding roughness elements to a base value for channel, bank, and vegetation characteristics. Estimates of roughness in the upper watershed utilized the work of Jarrett (1984) and Yochum, Bledsoe, David, and Wohl (2011), who developed empirical equations for steep environments. Roughness values derived by these four methods were employed in the Manning equation, and the resulting values of Q_{bkf} were compared with gage data and bankfull discharge estimates from regional curves (NRCS, 2012b). The resulting Q_{bkf} estimates were greater by an average of 5% than estimates from the regional curve at 8 of 10 sites, possibly due to orographic uplift at the escarpment (Lecce, 2000).

Representative particle sizes in metric units and phi units corresponding to specific percentiles (D_{95}, D_{90}, D_{75}, D_{84}, D_{50}, D_{25}, D_{16}, D_5) were calculated from the grain-size distributions (GSD). These percentiles are required for inclusive graphic statistics developed by Folk and Ward (1957). Coarse-bedded channels typically exhibit a bimodal GSD, but the fine-grained mode typically is not associated with the structural stability of the channel (Thompson et al., 2006; Wilcock, 2001). Therefore, percentiles were calculated after truncating the sample at 6 mm. Sample measurements of ≤6 mm were few, and truncation had little to no effect on median and upper percentiles used for hydraulic calculations.

The processed field data were analyzed first for downstream trends. DHG relationships were analyzed using traditional log–log power functions (Leopold & Maddock, 1953). Bed material calculations were analyzed for fining trends by fitting an exponential curve (Surian, 2002). Hydraulic variables, bed material, and bedform trends were analyzed together to understand potential controlling factors and quantitatively describe them through the transition zone.

Results and discussion

Substantial differences in morphology clearly occur in the downstream direction through the transition zone from the Blue Ridge to the Piedmont as shown by a high range of values in most calculations (Table 2). Results are presented in the following sections, first by the postulated downstream trends (DHG, bed material size, and bedforms) then by downstream trends in channel hydraulic parameters (e.g., shear stress, roughness measures). Finally, relationships between variables are presented to examine the nature of channel morphology and controls of downstream trends through the transition zone.

Downstream hydraulic geometry

Bankfull discharges (Q_{bkf}) computed from cross-section analysis can be expressed by a power function relationship with drainage area (Figure 3a). Width, depth, and velocity

Table 2. Summary of data and calculations. Variables are defined in Table 1.

			Bedform morphology				Grain size parameters			Cross sections	
	DA	*S*	*H*	*L*	*H/L*	*R/H*	D_{84} (mm)	D_{50} (mm)	R/D_{84}	*Area*	*WP*
max	110.8	0.0376	1.423	165	0.074	14.1	940	480	79.18	58.6	25.5
mean	N/A	0.0134	0.690	53.3	0.031	2.98	303	142	17.49	22.0	17.9
min	2.9	0.0003	0.163	4.40	0.003	0.612	17	10	1.01	4.24	9.43
Site 1	2.9	0.0326	0.450	13.500	0.033	0.998	293	123	1.53	4.2	9.4
2	12.7	0.0376	0.900	18.950	0.047	1.054	940	480	1.01	12.2	12.8
3	21.7	0.021	0.327	4.400	0.074	2.059	540	280	1.25	10.3	15.3
4	22.8	0.0198	0.600	9.100	0.066	1.205	655	235	1.10	10.7	14.7
5	31.5	0.006	1.423	26.750	0.053	0.612	145	72	6.00	20.1	23.0
6	62.5	0.007	1.230	165.00	0.007	0.856	232	136	4.54	20.0	19.0
7	79.2	0.0027	0.315	85.000	0.004	4.443	90	39	15.55	27.1	19.3
8	96.1	0.0003	0.163	62.850	0.003	14.130	17	10	135.07	58.6	25.5
9	103.9	0.0037	0.468	94.933	0.005	3.089	33	18.5	43.810	28.9	20.0
10	110.8	0.0035	1.020	52.800	0.019	1.396	68	32	20.936	27.8	19.5
		Cross-sections				Hydraulic Parameters					
Cont.	*R*	*w*	*d*	*f*	Q_{bkf}	*v*	τ	Ω	*W/D*	*ER*	*VW*
max	2.30	21.90	2.68	2.31	52.0	1.87	339	4944	23.3	6.00	500
mean	1.13	16.33	1.24	0.68	28.8	1.32	103	2013	14.8	2.61	153
min	0.45	8.86	0.48	0.09	2.99	0.71	6.75	132	8.19	1.48	20.0
Site 1	0.4	8.9	0.5	2.30	2.99	0.71	143.45	955.3	18.52	2.52	95
2	0.9	12.1	1.0	2.31	13.42	1.10	338.68	4943.9	12.03	1.48	20
3	0.7	13.4	0.8	0.57	14.31	1.39	138.58	2945.6	17.52	2.39	80
4	0.7	14.0	0.8	0.56	15.10	1.42	140.27	2929.9	18.33	2.78	250
5	0.9	21.6	0.9	0.40	20.23	1.01	51.19	1189.5	23.26	3.17	87
6	1.1	17.6	1.1	0.19	34.61	1.73	72.25	2374.4	15.43	2.23	300
7	1.4	17.7	1.5	0.13	40.33	1.49	37.03	1067.1	11.61	1.50	93
8	2.3	21.9	2.7	0.09	44.97	0.77	6.75	132.2	8.19	2.07	28
9	1.4	18.6	1.6	0.14	49.90	1.73	52.42	1809.4	11.97	1.94	75
10	1.4	17.6	1.6	0.11	52.04	1.87	48.83	1784.8	11.10	6.00	500

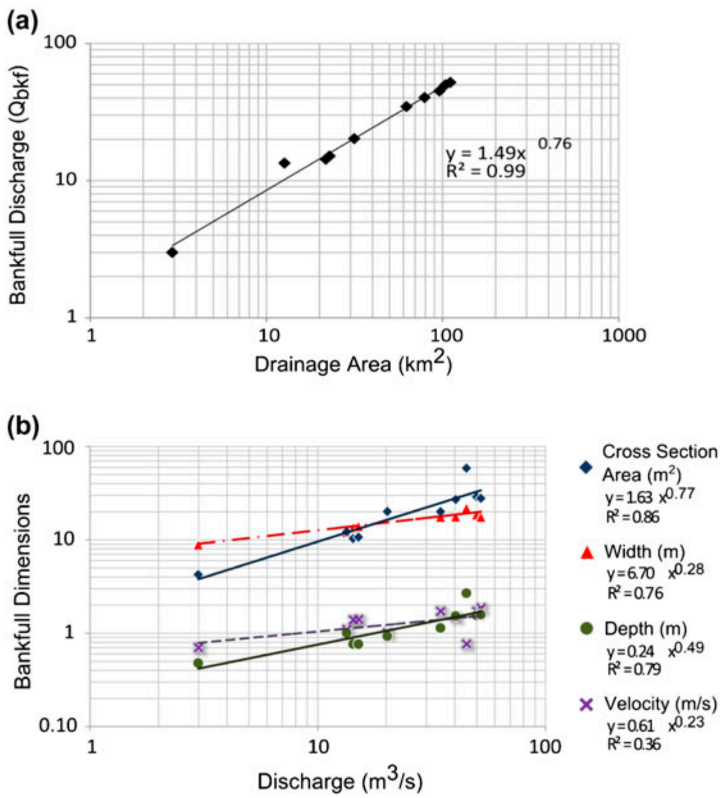

Figure 3. Downstream hydraulic geometry relationships among (a) discharge and drainage area and (b) discharge and top width, mean depth, cross-sectional area, and mean velocity. Middle Saluda River, South Carolina.

are also strongly correlated with bankfull discharge by log–log (power) functions throughout the study area (Figure 3b). The resulting R^2 values for width and depth exceed the threshold of 0.5 for well-developed DHG given by Wohl (2004). Well-expressed DHG relationships suggest that channels in the study area are adjusted to current sediment and discharge regimes at the scale of the study, from the steep, step-pool channels to the lower gradient pool-riffle channels.

A general DHG trend appears to exist in this basin, although the number of reaches in this study is limited and a fine-resolution analysis of channel geometry could result in weaker relationships (Fonstad & Marcus, 2010). One sample observation stands out as a high residual in all of the DHG models except for width. Although it is not treated as an outlier in this study, its removal from the model would improve the explained variance (R^2) substantially for depth and cross-sectional area (R^2 of 0.91 and 0.95, respectively). This point corresponds to a sudden and drastic decrease in slope (0.0003) caused by a resistant channel-spanning bedrock knickpoint; i.e., a dam-like effect influencing both hydraulics and sediment transport. The depth residual of this point is 2.5 times the standard deviation of depth residuals. Importantly, this point characterizes the signature of a forced morphology. The implications of such a reach for the dynamics of the transition zone are discussed in detail through the rest of the paper.

Bed material

Downstream fining of the 84th percentile (D_{84}) of channel-bed material follows an exponential decay trend with an R^2 of 0.74 (Figure 4). The fining coefficient of this model is 0.26 km^{-1}. Comparable coefficients have been reported to represent sorting processes in headwater and upper basin reaches (Surian, 2002). The uppermost sample location ('*x*' in Figure 4b) was not used to compute the curve because it represents a distinct geomorphic province above where the river plunges into the gorge downstream. Bed material size at this site is considerably smaller than at the downstream sites in and below the gorge. Downstream coarsening of headwater channels has been observed by Brummer and Montgomery (2003) in Washington and by Leigh (2010) in North Carolina.

Nine of 12 bed material samples were poorly sorted Folk and Ward (1957). The remainder were moderately well sorted (at sites 7 and 8) and moderately sorted (site 9). The largest bed material (D_{84} = 940 mm) was recorded at the reach with the steepest valley walls and highest gradient, which is the second downstream sample site. The finest D_{84} bed material size (17 mm) is located at the forced reach mentioned in the DHG analysis that is influenced by an abrupt structurally controlled break in channel slope. The channel bed in this reach consists mostly of sand with small patches of pebbles and fine gravel. The modal grain size in this area may be even finer than the D_{50} of this sample (10 mm), because the sample was taken from a single patch of gravels likely exposed by local scour.

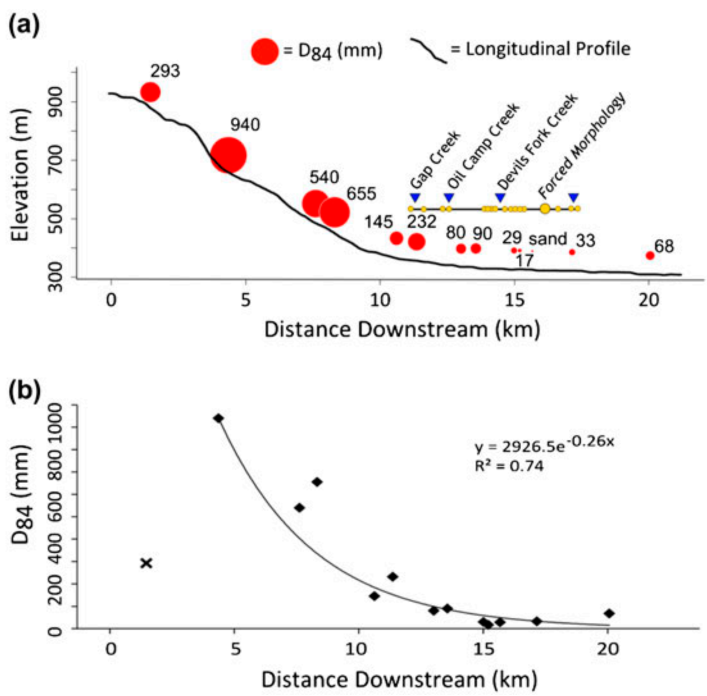

Figure 4. Downstream fining of D_{84} bed material in relation to longitudinal profile (a) with location of bedrock knickpoints and large tributaries and (b) expressed exponentially. Middle Saluda River, South Carolina.

Further, local fining occurs within the reach due to the damming effects of the resistant bedrock. The coarse fraction of bed material (D_{84}) decreases from 29 mm at the top of the reach to sand (≤2 mm) near the bedrock outcrop. Bed material caliber at this reach is representative of locally forced hydraulics rather than a systematic longitudinal continuum. As with the models of DHG, the general trend of downstream fining suggests systematic changes through the Blue Ridge-Piedmont transition. However, an increased number of samples, especially in forced hydraulic reaches, could reveal weaker trends and more complexity. Nonetheless, the downstream fining trend from the upper to lower reaches of the watershed serves as a foundation for relating bed material to bedforms.

Bedforms

Bedforms in the study area fit into three categories (*n* of each bedform type in parentheses): (4) step-pool, (5) pool-riffle, (1) and infilled (forced) morphology. These categories were determined in the field by comparison with photographs and physical descriptions of each bedform type as discussed in the literature (i.e., Montgomery & Buffington, 1997; Thompson et al., 2006; Wohl & Merritt, 2008). The uppermost four reaches were characterized as step-pool morphology, although differentiation between cascade and step-pool bedform types is not always clear. Thompson et al. (2006) described an intermediate bedform (*cascade-pool*) that is to some degree a combination of the two. Although it is recognized that this category may be an appropriate description for the bedforms encountered, there are not enough data in this study to warrant dividing the step-pool morphologies into intermediate morphologies. Further, it suffices to use the term *step-pool* here because the dominant contrast in bedform types occurs downstream where pool-riffles and forced morphologies commence. No cascades were recognized in the study reaches. They may be present in the watershed, but step-pool morphologies dominate.

The spatial pattern of channel bedforms generally follows the typical downstream progression outlined in Montgomery and Buffington (1997), with step-pool reaches giving way to pool-riffle morphologies. No plane-bed channels were observed in the short downstream transition between step-pool and pool-riffle channel types, and it is possible that there are no plane-bed reaches, as observed by Thompson et al. (2006) in some basins with granite lithology. The infilled morphology is situated longitudinally between pool-riffle channels. It is forced by a local gradient decrease due to a bedrock knickpoint (see DHG above) and characterized by a nearly featureless sand bed (see *Bed Material* above).

Relating hydraulics, sediment, and bedforms

The transition between bedform types coincides with the systematic decrease in bed material size and increase in DHG variables. However, these general trends can be artifacts of basin-scale changes explained by hydraulic processes that vary at a finer scale. An analysis of hydraulic variables, sediment, and bedforms through the transition zone reveals relationships that could be driven by more local factors. Cross-sectional stream power (Ω) and mean boundary shear stress (τ) in the watershed generally decrease downstream, as expected. However, they also follow closely to more complex variations in bed material that are superimposed on the downstream fining trend. Slope, τ, and Ω often explain variance in bed material size that is not explained simply by progressive downstream fining with drainage area, especially in mountain drainages with irregular longitudinal profiles and forced morphologies (Thompson et al., 2006). In this case, D_{84} and D_{50} are strongly correlated with τ (r = 0.93 and 0.95), Ω (r = 0.89 and 0.91), and slope (r = 0.85 and 0.83), which suggests a high degree of hydraulic influence on the size of exposed bed material relative to sediment inputs. The infilled morphology has the finest bed material (D_{50} = 10) and the lowest S (0.0003), τ (6.75), and Ω (132), despite the greatest cross-section area and d and an intermediate drainage area of 96 km^2 and Q_{bkf} of 45 $m^3 s^{-1}$.

Comparisons between studies reveal an overlap in the ranges of *individual* values of slope, grain size, and drainage areas for bedform types, but the combination of these variables has been shown to distinguish bedform types in different parts of the world (Montgomery & Buffington, 1997; Thompson et al., 2006; Wohl & Merritt, 2008). A sediment regime diagram can be used to distinguish among reach-scale bedform types by applying dimensionless surrogates for the Montgomery and Buffington (1997) variables of sediment supply and transport capacity (Thompson et al., 2006). Dimensionless bedload transport (q_b*), a surrogate for sediment supply, is estimated by:

$$q_b^* = 8(\tau^* - \tau_{c50}^*)^{1.5} \tag{1}$$

where τ* is bankfull shear stress (Shields), defined as:

$$d^* S / 1.65 D_{50} \tag{2}$$

where and D_{50} is median particle size, d is mean bankfull depth, and 1.65 is the submerged specific gravity of sediment. τ^*_{c50} is dimensionless critical stress of D_{50}, which is set at 0.03 (Buffington, Woodsmith, Booth, & Montgomery, 2003).

Dimensionless discharge per unit width (q*), a surrogate for transport capacity, is:

$$q^* = \frac{ud}{(1.65\, gD_{50})^{0.5} D_{50}} \tag{3}$$

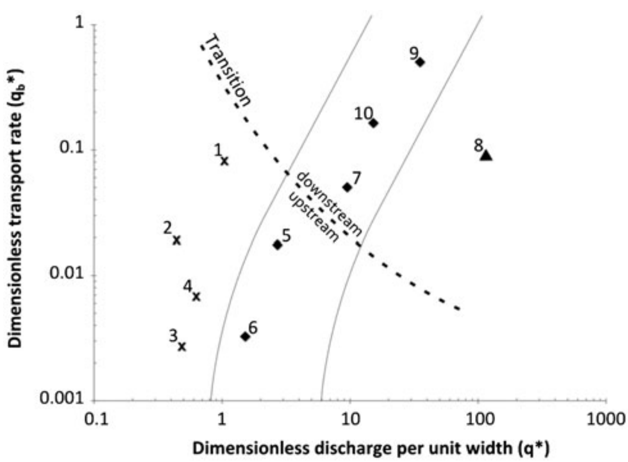

Figure 5. Middle Saluda River data plotted on a sediment regime diagram as utilized by Thompson et al. (2006) for mountain streams. Lines represent natural divisions in data separating bedform types: 'x's are step-pool, diamonds are pool-riffle, and the triangle is infilled. The dashed transition line refers to the key transition zone in the discussion.

where u is vertically averaged velocity and g is acceleration due to gravity.

As applied to the data in this study, the plot of q_b^* and q^* distinguishes reach morphologies (Figure 5) independently of drainage area (scale). The numbers on Figure 5 represent the downstream order of reaches. The forced morphology at site 8 is distinguished from the rest. Moreover, sites 5 and 6 are distinguished from the rest of the pool-riffle types because of the coarser associated bed caliber and steeper slopes, which may suggest a different pool-riffle regime than downstream. Site 1 is also correctly classified as step-pool regardless of its exclusion from the downstream fining trend. This suggests that sediment regime analysis could be a useful tool in a basin with a series of forced morphologies as it incorporates a combination of the fundamental controls of bed morphology and may predict variations in morphology types in basins where slope-area relationships are not well developed. Carbonneau et al. (2012) suggested that local variability that is not explained simply by downstream trends (scale) can be a dominant feature in some fluvial systems.

Downstream bedform transition

Bedforms can be interpreted in terms of roughness and channel resistance. With a deformable boundary under certain discharge and sediment supply conditions, channel resistance is maximized by the topography of bedforms (Montgomery & Buffington, 1997; Wohl & Merritt, 2008). *Relative grain submergence* (R/D_{84}) increases downstream, i.e., grains protrude into a smaller proportion of the flow and cause less resistance. Similarly, bedforms generate roughness that can be expressed by *relative form submergence (R/H)*. Wohl and Merritt (2008) present plots of Darcy–Weisbach friction factor (f) versus R/D_{84} that show no differences between values of f in pool-riffle vs. step-pool channel types in response to increasing in R/D_{84}. This suggests that bedform roughness may compensate for decreasing grain roughness (i.e., increasing R/D_{84}) downstream in mountain environments. The vertical height (H) or amplitude of bedforms adjusts to hydraulic variables (decreased slope, increased depth), so that R/H tends to be conservative, ultimately minimizing the variance of hydraulic roughness in the downstream continuum from step-pool to pool-riffle channels (Wohl & Merritt, 2008).

In contrast with the above theory, the analysis of f and R/D_{84} in this study reveals a statistically significant (F = 20.78, p = 0.00186) decreasing trend in f (Figure 6a). This trend is consistent with trends observed in lower gradient channels (Knighton, 1998). The sequence of sites in this plot is generally consistent with a downstream trend except for the infilled site (site 8). Plots of R/D_{84} and R/H also show a systematic trend; R/H and R/D_{84} increase together (Figure 6b), which is contrary to the hypothesis that R/H should remain consistent as R/D_{84} increases. Figure 6 suggests that downstream bedforms are not completely compensating for decreasing grain roughness. Rather, drastic decreases in gradient associated with the transition zone may be influencing downstream hydraulics so that complete compensation of bedform dimensions is unnecessary to achieve minimum variance in resistance.

The spatial locations of points in Figure 6 indicate that R/D_{84} and R/H are not simply related to drainage area. The hydraulic controls are associated with location relative to landscape features. The three points with the highest R/H values (9, 7, and 8 in Figure 6b) are located in a part of the watershed where gradient is structurally controlled by the presence of erosion-resistant bedrock knickpoints. Thompson et al. (2006) hypothesized that subtle "macro-scale" (i.e., broader than reach-scale) features

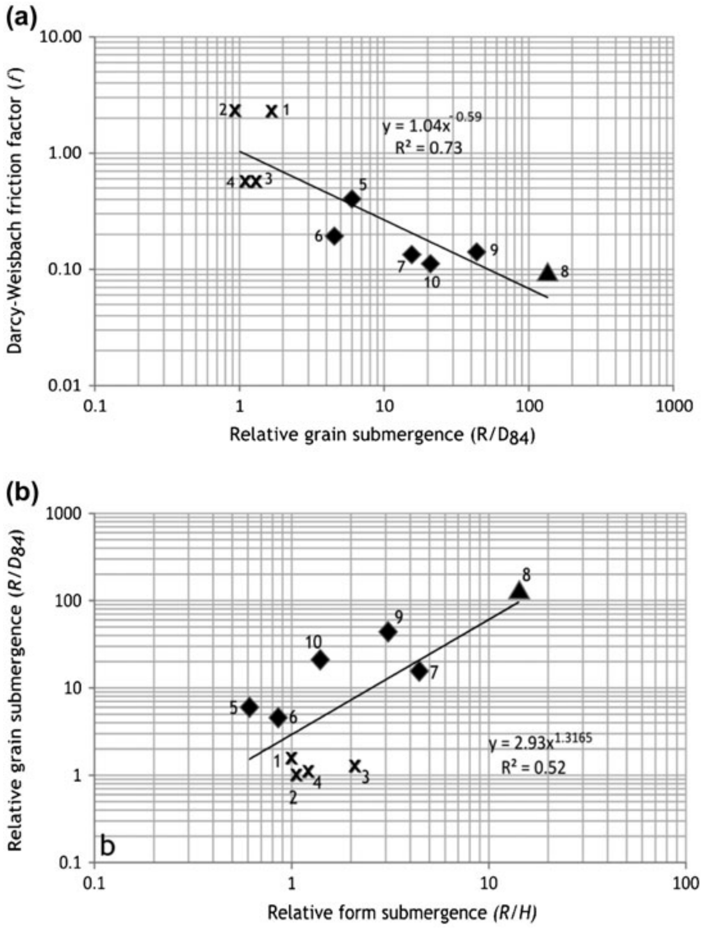

Figure 6. Middle Saluda River roughness data, plotted as (a) Darcy–Weisbach friction factor versus R/D_{84} and (b) Relative grain submergence (R/D_{84}) versus Relative form submerge (R/H). Numbers are downstream order of sample reaches. Bedform type symbols are as in Figure 5.

influenced the large number of intermediate channel morphologies in their study. Myers and Swanson (1997) suggested similar gradient influences on pool-to-pool spacing in forest streams. It is hypothesized here that decreases in gradient associated with longitudinal steps (i.e., knickpoints) are influential to bedform dimensions, hence the trends noted in Figure 6. This is partially in conflict with the assumption of DHG that bedforms in mountain rivers adjust in proportion to drainage area (scale), as in completely alluvial-controlled longitudinal profiles. Similarly, the same three reaches (sites 7, 8, and 9; one infilled and two pool-riffle) have the smallest bedform ratios (H/L) and drive the increasing trend in Figure 6b. It appears that the trends in Figure 6 indicate forced gradients that govern the transition of hydraulic processes to another regime.

Key transition zone

One interpretation of the results of this study is that a key transition zone from the steep Blue Ridge into the Piedmont can be defined where characteristics of the channel change substantially. This zone is influenced by bedrock knickpoints as the river crosses structural ridges; substantial tributary inputs, one of which (Gap Creek) nearly doubles the drainage area; substantial decreases in long profile slope; and increases in sand inputs from the basin. Steep major tributaries meet the main channel near the inflection point of slopes (Figures 4, 5), causing substantial changes in discharge and sediment supply (particularly sand) over a relatively short distance at a point where slope is forced to decrease dramatically.

The data support the concept of a key transition zone in this watershed. The threshold of 10,000 kg s^{-3} (Wohl, 2004) is located at the beginning of the transition zone at (site 5), just downstream of the confluence with Gap Creek. Further, *w*/*d* peaks before the transition zone (site 3), then decreases, and is similar at the last four sites. The watershed could also be characterized by two separate bed material zones, one upstream and one downstream of the transition (Figure 4). Although the shift of bedforms from step-pool to pool-riffle occurs at a line above the transition zone, the sediment regime diagram (Figure 5) discerns pool-riffles upstream and downstream of the transition zone (sites 5 and 6 vs. 7–10). The lack of observed plane-bed channel types may indicate the abrupt nature of the transition. Further, the trends in flow resistance measured by f, R/D_{84}, and R/H (Figure 6) are also strongly influenced by the sample sites in the transition zone. Moreover, accumulation of sand in the Middle Saluda River from upstream tributaries has the potential to overwhelm bedforms where slope drastically decreases. Probed sand depths in pools reveals increased storage of sand after the transition zone, from an average of 15 cm at site 5 to >90 cm at sites 8 through 10 (although the areal extent of sand covering the channel bed is much greater at site 8). This suggests that sand is increasingly available to influence forced morphologies downstream.

Slope is a critical factor in the transition zone, and it is driven by external controls at a location in the landscape that accentuates the transition of bedforms. The key transition zone consists of 6 km of river punctuated by 17 erosion-resistant bedrock knickpoints, with a central zone of greater density (2 km; n = 11) where the river cuts across a substantial ridge. While the influence of individual bedrock knickpoints is easily recognized in forced morphologies at a specific stream reach (i.e., site 8), the structural control as a whole may be broadly influential in that it can be viewed as a local base level control (Ferguson & Ashworth, 1991; Fryirs et al., 2007). Thus, sites 7 and 9, inside this structurally controlled zone, may be regarded as subtly forced morphologies that are controlled by base level control of knickpoints, although their morphological changes are not drastic. This can extend the concept of "forced" beyond the local reach-scale to the "macro-scale" (Thompson et al., 2006).

Conclusion

Well-defined trends in DHG and bed material caliber for the watershed do not fully reflect the variability or the complex nature of the transition suggested by analysis of scale-independent hydraulic factors and reach-scale bedform morphologies. DHG, bed material caliber, and bedforms follow a general progression of upper step-pool-type morphologies to pool-riffle-type morphologies downstream. However, rather than a simple gradual progression, as suggested by DHG and downstream fining models, this analysis suggests a clear transition zone punctuated by forced morphologies at reach and segment scales. This clear transition is hypothesized to be a result of a break-in-slope associated with major tributary confluences and a stepped longitudinal profile (e.g., bedrock knickpoints) through the transition zone. Bedforms at longitudinal steps may be anomalous with regard to drainage area, but they are predictable by hydraulic and sedimentological factors. A better understanding of channel morphology will come from an understanding of watershed features that govern these local hydraulic and sedimentological factors, such as changes at geologic structures and tributary junctions.

The findings of this study suggest that DHG and downstream fining models should only be viewed as general trends and that interpolation of fluvial characteristics from these generalizations should be applied with caution. Analysis independent of drainage area should be utilized and placed in a landscape context when using models of morphology for restoration or management objectives in mountain environments.

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