

Lithological and fluvial controls on the geomorphology of tropical montane stream channels in Puerto Rico

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ABSTRACT: An extensive survey and topographic analysis of five watersheds draining the Luquillo Mountains in north-eastern Puerto Rico was conducted to decouple the relative influences of lithologic and hydraulic forces in shaping the morphology of tropical montane stream channels. The Luquillo Mountains are a steep landscape composed of volcaniclastic and igneous rocks that exert a localized lithologic influence on the stream channels. However, the stream channels also experience strong hydraulic forcing due to high unit discharge in the humid rainforest environment. GIS-based topographic analysis was used to examine channel profiles, and survey data were used to analyze downstream changes in channel geometry, grain sizes, stream power, and shear stresses. Results indicate that the longitudinal profiles are generally well graded but have concavities that reflect the influence of multiple rock types and colluvial-alluvial transitions. Non-fluvial processes, such as landslides, deliver coarse boulder-sized sediment to the channels and may locally determine channel gradient and geometry. Median grain size is strongly related to drainage area and slope, and coarsens in the headwaters before fining in the downstream reaches; a pattern associated with a mid-basin transition between colluvial and fluvial processes. Downstream hydraulic geometry relationships between discharge, width and velocity (although not depth) are well developed for all watersheds. Stream power displays a mid-basin maximum in all basins, although the ratio of stream power to coarse grain size (indicative of hydraulic forcing) increases downstream. Excess dimensionless shear stress at bankfull flow wavers around the threshold for sediment mobility of the median grain size, and does not vary systematically with bankfull discharge; a common characteristic in self-forming 'threshold' alluvial channels. The results suggest that although there is apparent bedrock and lithologic control on local reach-scale channel morphology, strong fluvial forces acting over time have been sufficient to override boundary resistance and give rise to systematic basin-scale patterns. Copyright © 2010 John Wiley and Sons, Ltd.

KEYWORDS: lithology; hydraulics; morphology; mountain streams; channel profiles; colluvial–fluvial processes

1. Introduction

Recent advances in understanding the linkages between tectonics and surface processes have spurred interest in the evolution of mountain and bedrock streams (Whipple, 2004; Bishop, 2007). Mountain stream channels have complex morphologies and a number of studies implicate several different controls on their development, including tectonic and structural (VanLaningham *et al.*, 2006), bedrock (Snyder *et al.*, 2003), storm pulses (Gupta, 1988), and non-fluvial processes such as landslides/debris flows (Brummer and Montgomery, 2003; Stock and Dietrich, 2006) and glaciers (Wohl *et al.*, 2004). Other studies have demonstrated the characteristic morphology of mountain streams (Grant *et al.*, 1990; Montgomery and Buffington, 1997; Wohl and Merritt, 2001), their hydraulic geometry (Wohl, 2004), the complexity of

sediment transport (Blizard and Wohl, 1998; Lenzi *et al.*, 2004; Torizzo and Pitlick, 2004), and the distribution of sediment within mountain drainages (McPherson, 1971; Grimm *et al.*, 1995; Pizzuto, 1995; Constantine *et al.*, 2003; Golden and Spring, 2006). Montane streams in the tropics are among the most extreme fluvial environments in the world (Gupta, 1988). A combination of steep slopes, high mean annual rainfall, and intense tropical storms generate an energetic and powerful flow regime. The high rates of erosion and dramatically dissected landscapes prevalent in the world's tropical mountainous regions attest to the erosive power of these rivers. Yet the channel morphology that is sculpted by fluvial and non-fluvial processes in tropical montane environments is generally unknown. This paper investigates controls on mountain stream channel morphology in the Luquillo Mountains of Puerto Rico, a tectonically active landscape with

varying bedrock and structural resistance that is rapidly eroding due to extremely wet tropical conditions, frequent intense storms, and a high susceptibility to mass-wasting.

Montane streams in both tropical and temperate environments share some common characteristics. A combination of active tectonic uplift and resistant lithologies that are common in many mountainous regions yield steep-gradient channels that are dominated by bedrock and coarse clasts (Grant *et al.*, 1990). Vertical valley walls and confined channel boundaries inhibit floodplain development and may locally determine channel width (Montgomery and Gran, 2001; Finnegan *et al.*, 2005). Longitudinal profiles are typically segmented by knickpoints and waterfalls. Furthermore, montane streams often have high boundary roughness, intense hydraulic turbulence, high entrainment rates and stochastic bedload movement (Wohl *et al.*, 2004).

However, some tropical mountain streams may have unique features that vary from their temperate counterparts. The absence of glaciation excludes glacial landforms, such as u-shaped valleys and coarse moraine deposits, that are prevalent in some temperate montane basins. Relatively high rates of chemical and physical weathering rapidly denude tropical landscapes and may affect rates of channel-sediment diminution and patterns of downstream fining (Brown *et al.*, 1995; White *et al.*, 1998; Rengers and Wohl, 2007). Frequent landslides triggered by heavy rains introduce pulses of coarse sediment to the channels and strongly link fluvial and colluvial forces (Larsen *et al.*, 1999). Large woody debris that is common in temperate streams is rapidly decomposed in the tropics, despite high inputs from surrounding mature forests and hurricanes (Covich and Cowl, 1990). Rapid runoff production generates flashy, frequent, short-duration floods (Schellekens *et al.*, 2004; Niedzialek and Ogden, 2005), and occasional high-magnitude floods associated with hurricanes and other tropical disturbances effectively rework boulder channels (Gupta, 1975; Scatena and Larsen, 1991).

The relatively few studies that have addressed the underlying controls structuring the morphology of tropical mountain streams demonstrate the influence of a variety of factors. Ahmad *et al.* (1993) and Gupta (1995) concluded that the lithology of many streams in the Caribbean plays a strong role in locally determining channel morphology, dictating the course of the river, and governing the distribution of large boulders. These streams commonly have bedrock and boulder-lined channels, whereas traditional depositional forms built by sand, gravels, and cobbles are sparsely observed. In some streams in Jamaica and Puerto Rico, Gupta (1975) emphasized the role of high discharge relative to drainage area as a key hydraulic control shaping channel morphology. Lewis (1969) demonstrated that local lithologic factors, such as bed material cohesion and channel constriction, influenced at-station hydraulic geometry of a river network in north-central Puerto Rico. Yet it was also demonstrated that consistent scaling of downstream hydraulic geometry was developed across multiple lithologies. Similar characteristics were noted in the Río Chagres in Panama, where hydraulic controls due to notably high unit discharge are apparently sufficient to override lithologic controls and develop a river network with well-developed downstream hydraulic geometry (Wohl, 2005). This study also utilized extensive basin-scale field reconnaissance to quantify downstream patterns in tropical mountain stream morphology, and contended that similar high-quality field surveys in different tropical regions are essential to further knowledge about these systems.

Geomorphic processes that structure stream networks can be inferred from the topography of the drainage basin and from patterns in channel characteristics that can be measured

in the field. Measures for quantifying and comparing stream geomorphic processes at the basin-scale include: longitudinal profiles, hydraulic geometry, grain sizes, and the spatial distribution of fluvial energy expenditure and shear stresses. The downstream and spatial patterns of these measures can indicate underlying controls on channel geomorphology. For example, many alluvial rivers develop systematic changes in slope, channel geometry, and grain size from their headwaters to the coast in response to changes in discharge and sediment yield (Paola and Seal, 1995). These changes result in many well-known basin-scale patterns, such as concave-upward longitudinal profiles, well-developed hydraulic geometry, and progressive downstream fining of sediment. Conclusions about the relative importance of lithologic and hydraulic controls on stream network development can be drawn from a basin's adherence to, or significant deviations from, these theoretical geomorphic patterns.

Longitudinal profiles of rivers often reflect the lithologic and tectonic controls on channel development (Kirby and Whipple, 2001). A theoretical profile of a graded stream has a smoothly concave-upward shape; steep in the headwaters and flat near the mouth (Hack, 1957). A river of this form has achieved an assumed balance between the erosion from fluvial processes and the resistance from lithologic and tectonic forces. Deviations from this idealized grade, such as changes in concavity (Seidl *et al.*, 1994) and the presence of segmentation/knickpoints (Crosby and Whipple, 2006; Goldrick and Bishop, 2007) can indicate the influence of non-fluvial forces. For example, faults and structural barriers may confine a river and constrain slope (Whipple, 2004), multiple bedrock units with varying resistance to river incision often create slope breaks, protrusions, and/or knickpoints (Wohl *et al.*, 1994), and landslides/debris flows can locally constrain the channel gradient and concavity. (Grant *et al.*, 1990; Stock and Dietrich, 2006).

Downstream hydraulic geometry (DHG) characterizes systematic downstream changes in channel geometry as power-law relationships with discharge, and may be used to quantify the influence of fluvial controls on channel form (Leopold and Maddock, 1953). DHG has successfully described river patterns worldwide in many physiographic environments. However, it is intended to describe changes in self-formed alluvial rivers that readily adjust their geometry in response to changes in discharge and sediment transport. The ubiquity of well-developed DHG in these self-formed rivers has been explained from a combination of basic hydraulic and sediment transport processes (Singh, 2003; Parker *et al.*, 2007). However, the complicated hydraulics and sediment transport processes associated with boulder- and bedrock-armored channels in many mountain rivers may confound these relationships. Consistent power-law relations in downstream channel geometry have been observed in some mountain rivers, even though these streams alter their morphology at longer time scales than most alluvial rivers (Molnar and Ramirez, 2002; Wohl and Wilcox, 2005). In fact, mountain rivers with well-developed DHG tend to have an above-average ratio of total stream power (a measure of hydraulic driving forces) to coarse grain size (a measure of boundary resistance) (Wohl, 2004). In contrast, mountain rivers that are strongly controlled by geologic rather than hydraulic controls will often display poorly-defined DHG (Wohl *et al.*, 2004).

The distribution of grain sizes throughout the stream network can also yield insight into the underlying lithologic controls. Grain size in the stream channel is largely dependent on the underlying bedrock, the input from hillslopes, and the mechanisms of weathering and transporting clasts. In fluvial systems where the bed material is readily mobile, there is often a balance between discharge, sediment transport, and slope.

Consequently, grain size often declines with increasing drainage area such that the largest grains are found in headwater channels and smaller grains are found in lower reaches (Paola and Seal, 1995; Rice, 1999; Constantine *et al.*, 2003). In steep montane catchments where landslides introduce large pulses of coarse (and potentially immobile) sediment, this balance is upset, and grain-size patterns are often discontinuous (Grimm *et al.*, 1995; Pizzuto, 1995; Brummer and Montgomery, 2003). The downstream pattern in grain sizes indicates the relative influence of coarse material deposited from hillslope processes and the ability of the channel to transport sediment.

Many river networks also tend towards an assumed optimal state of energy expenditure throughout their evolution such that certain indices of energy expenditure are either constant or linear along the river profile (Molnar and Ramirez, 2002). Nonlinearity in stream power, whereby energy expenditure is concentrated in specific reaches rather than uniformly dispersed, can indicate underlying geologic control (Graf, 1983; Lecce, 1997). Similarly, many stream networks have a mid-basin maximum in stream power, the location of which is dependent on slope, the flow regime, and the structure of the basin (Knighton, 1999). Large gradients in bed stress or energy expenditure also yield gradients in sediment flux, causing certain parts of the river to erode and others to deposit sediments in an effort to remove these gradients. In bedrock- and boulder-lined channels where coarse sediment is not readily mobile, the ability of the channels to adjust their morphology to remove these gradients in energy expenditure may be hindered.

Lastly, the downstream trend in boundary shear stress at bankfull discharge provides insight into sediment mobility and the relative stability of channels. The Shields parameter, a dimensionless bed shear stress that is expressed as a ratio of slope, depth, and size of the bed material, is a quantitative indicator of flow competence and is strongly related to alluvial channel form (Dade and Friend, 1998; Dade, 2000). In many self-forming alluvial channels, the Shields parameter at bankfull flow does not vary systematically throughout a basin. Assuming a constant critical dimensionless shear stress, this lack of scaling between the Shields stress and bankfull discharge implies that many alluvial channels are at the threshold for incipient sediment mobility. However, in gravel- and boulder-lined mountain channels, both the Shields parameter and critical dimensionless shear stress often vary widely throughout the basin, depending on flow resistance (Mueller *et al.*, 2005). If the flow regime in montane channels is sufficient to mobilize the bulk of the sediment, we would expect the Shields parameter to be consistently higher than the critical dimensionless shear stress; lower if the sediment is too coarse to transport. Furthermore, if the controls on sediment transport shift from non-fluvial forces upstream to fluvial forces downstream, we would expect the excess dimensionless shear stress to increase with bankfull discharge.

In this study, we address the need for research on the geomorphology of tropical mountain rivers by quantifying basin-scale geomorphic patterns and processes in five adjacent watersheds in the Luquillo Mountains of northeastern Puerto Rico. Using data from an extensive field survey of the stream networks, coupled with GIS-based topographic analysis, we compare channel profiles and subsequent downstream changes in cross-sectional geometry, grain sizes, stream power, and shear stresses. This comprehensive dataset allows us to test several competing hypotheses regarding the controls on the stream channel morphology.

First, local lithologic factors may solely structure the form of the river. If these lithologic influences (varying strength of different rock types, resistant bedrock channel boundaries,

and coarse immobile sediment delivered to the channel by landslides) exclusively sculpt the geometry of the channel and determine sediment characteristics, we would expect to see morphologic trends that strongly deviate from idealized downstream patterns. Indicators of strong lithologic control would include segmented channel profiles, poorly developed hydraulic geometry, random downstream patterns in grain size, and insufficient stream power and boundary shear stress to mobilize available sediment.

Second, hydrologic forces may overcome lithologic resistance to shape the morphology of the channel. If the high unit discharge and associated stream power of the energetic tropical flow regime are sufficient to overcome lithologic resistance and mobilize coarse sediment, then we would expect downstream changes in channel morphology to have systematic trends similar to those found in many fully alluvial rivers. Specifically, indicators of hydrologic control would include smoothly graded profiles, well-developed downstream hydraulic geometry, progressive downstream grain size fining, and constant indices of stream power and shear stress throughout the basin.

Third, a combination of both lithologic and hydrologic forces may act upon the channel, but their relative control on the morphology varies throughout the basin. In this case, it is likely that lithologic factors control the channel morphology in the montane reaches where colluvial and bedrock processes dominate, followed by a transition to hydrologic control in lower reaches where the river leaves the mountains and enters the coastal plain. Evidence for this hypothesis would include a mid-basin transition in channel profile concavity, the degree that downstream hydraulic geometry is developed, grain size coarsening/fining, and a change in sediment mobility. Such patterns would suggest that the relative magnitude of colluvial and fluvial forces varies spatially and that both types of forces structure the morphology of the river to differing degrees at varying scales.

Study Area

The Luquillo Mountains in north-eastern Puerto Rico rise steeply from sea-level to over 1000 m in elevation over a distance of 15 to 20 km. They are characterized by steep slopes, rugged peaks, and highly dissected valleys. The landscape is composed of several lithologies and a variety of land cover. The streams have their headwaters in the Luquillo Experimental Forest (LEF), a 113 km² protected forest reserve under the management of the United States Forest Service. The study area consists of five adjacent watersheds draining the LEF: Río Blanco, Río Espiritu Santo, Río Fajardo, Río Mameyes, and Río Sabana (Figure 1). The watersheds are similar physiographically, although they vary in size, lithology, and land cover. Drainage areas of these watersheds are 72 km², 92 km², 67 km², 44 km², and 35 km², respectively. All of the watersheds, except for the Río Sabana, reach the upper-most ridges of the Luquillo Mountains.

The humid subtropical maritime climate is influenced by both north-easterly trade winds and local orographic effects that interact to cause steep gradients in precipitation. Mean annual rainfall increases with elevation from approximately 1500 mm yr⁻¹ at the coast to >4500 mm yr⁻¹ at elevations above 1000 m (García-Martinó *et al.*, 1996). The principal weather systems affecting climate are convective storms, easterly waves, cold fronts, and tropical storms (van der Molen, 2002). Rainfall is a near-daily occurrence (Schellekens *et al.*, 1999), and high-intensity rainfall events and floods can occur in any given month. Hurricanes and tropical storms are

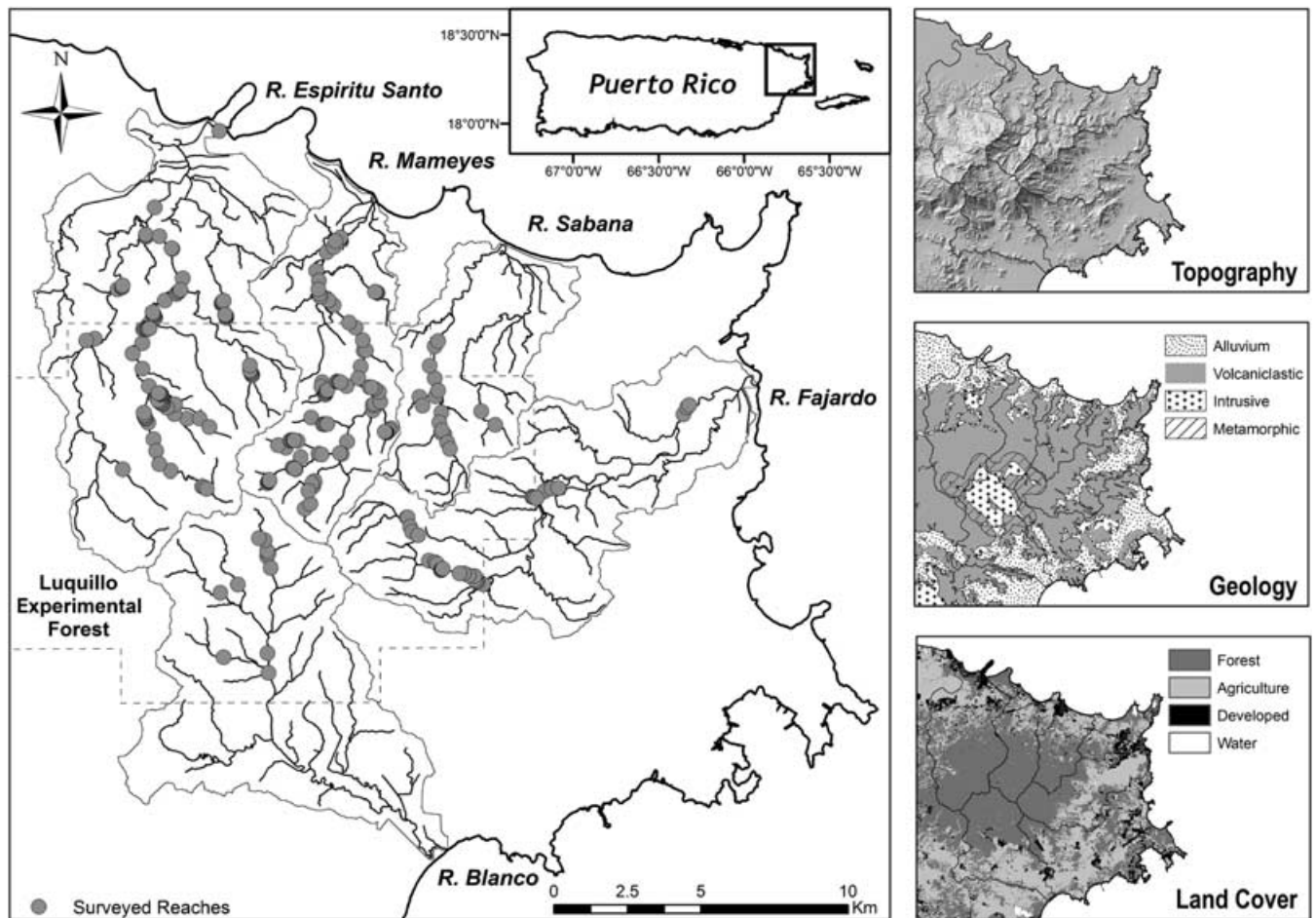


Figure 1. Location map of northeastern Puerto Rico. Shown are the 238 surveyed reaches in five adjacent watersheds and the regional topography, geology, and land cover.

common from August through October, and typically bring high daily rainfall in excess of 200 mm day^{-1} (Heartsill-Scalley *et al.*, 2007); the maximum recorded daily rainfall is $>600 \text{ mm/day}$ (Scatena and Larsen, 1991).

The streams of the Luquillo Mountains have been classified as 'flood dominated' channels that have a hydrologic setting similar to other montane environments in the Greater Antilles and regions along active tectonic zones in the humid tropics (Gupta, 1988; Ahmad *et al.*, 1993). Floods are intense and peak discharges can be 1000 times greater than baseflow. The unit discharge at baseflow is approximately $0.02 \text{ m}^3 \text{ s}^{-1} \text{ km}^{-2}$, whereas the highest peak unit discharge ever recorded at a regional stream gage was $19.7 \text{ m}^3 \text{ s}^{-1} \text{ km}^{-2}$ (United States Geological Survey, updated 2006). Peak-flow hydrographs are short-lived and typically have a duration of less than 1 h. Stormflow runoff is quickly flushed through the system such that the streams return to baseflow within 24 h of large events. Large floods are driven by storm events, as opposed to the seasonal floods associated with snowmelt in many temperate mountain streams. Consequently, discharges that are close to the annual peak are often experienced independently several times in a year (Scatena *et al.*, 2004).

The Luquillo Mountains were formed by early Tertiary volcanism and tectonic uplift associated with oceanic island-arc subduction. The landscape consists of several dominant lithologies: volcaniclastics, plutonic intrusions and dikes, contact metamorphics, and alluvium (Seiders, 1971a; Briggs and Angular-Cortés, 1980). The volcaniclastic rocks, comprised of marine-deposited volcanic sediments of late Cretaceous age, form the bulk of the Luquillo Mountains. They include

units of sandstones, siltstone, mudstones, breccias, conglomerates, tuff, and lava, that are complexly faulted and steeply tilted ($>30^\circ$). A Tertiary quartz diorite (granodiorite) batholith underlies the southern side of the study area. It outcrops in an area of approximately 24 km^2 , and is drained largely by the Río Blanco watershed, but also by small parts of the upper Río Espiritu Santo and Río Mameyes watersheds. It is rapidly eroding at an estimated denudation rate of 25–50 m per million years; one of the highest documented weathering rates of silicate rocks on the Earth's surface (Brown *et al.*, 1995; White *et al.*, 1998). A 1–2 km zone of contact metamorphism surrounds the granodiorite. These contact metamorphosed volcaniclastic rocks (hornfels facies) exhibit greater hardness than both their unmetamorphosed equivalents and the granodiorite (Seiders, 1971b). Because of their relative resistance to erosion, these rocks form steep cliffs and the tallest peaks in the region. Several vertical dikes traverse the volcaniclastic rocks, mainly at lower elevations ($<150 \text{ m}$). The mountains are fringed by a lowland coastal plain composed of Quaternary alluvium.

Past climates in the region are thought to be similar to the present, due to both comparable elevation of the mountains and the location of the mountain range in the subtropical maritime belt (Graham and Jarzen, 1969). Pollen assemblages and plant microfossils of subtropical and warm-temperate communities found in sedimentary sequences on the island suggest that the mountains of Puerto Rico had a comparable climate in the Oligocene. Modern climate evidence from a paleosol in the region indicates that the Luquillo Mountains' climate is generally considered to have oscillated around a

humid subtropical state throughout the late Pleistocene, without glaciers or dramatic changes (Scatena, 1998).

The riverine landscape is thought to be relatively old – on the order of several to tens of millions of years – based on both the plutonic and stratigraphic history. The emplacement of the plutonic rocks and supposed uplift of the Luquillo fault block occurred in the Eocene (Cox *et al.*, 1977; Kesler and Sutter, 1979), although significant erosion of the landscape and initial formation of the modern stream network probably did not occur until the Miocene (Monroe, 1980). Stratigraphic records preserved on the north-western side of the island suggest that surface streams began delivering clastic sediment into the sea during this time period.

All five watersheds currently drain protected primary forest in their upper elevations, mature secondary forest at mid-elevations, and both abandoned (reforesting) agricultural fields/grazing pastures and scattered urbanized developments along the coastal plain. Each river flows through a mangrove-lined estuary before reaching the coast. However, the land cover in the region has been continually changing since the Spanish colonization of the island in the late 17th century. Many low-elevation areas (<300 m) of north-eastern Puerto Rico were cleared for agriculture between 1830 and 1950. This caused an estimated 50% increase in runoff, and an order of magnitude increase in sedimentation on the coastal plain (Clark and Wilcock, 2000). Subsequent land clearance on steep valley slopes resulted in widespread erosion and landslides that delivered a large load of coarse sediments to the river. Large portions in the upper elevations of the LEF were never deforested during the 19th and 20th centuries due to government protection, steep slopes, and high rainfall (Scatena, 1989). Since 1950, urbanization and reforestation of former agricultural land in low-lying areas has resulted in elevated storm runoff and decreased sedimentation, allowing transport of previously deposited coarse alluvial sediment in coastal plain streams (Clark and Wilcock, 2000; Wu *et al.*, 2007).

Hillslopes are typically steep, in excess of 30° in many headwater areas, and are consequently strongly linked to channel processes (Scatena and Lugo, 1995). Landslides are the dominant process that physically weathers the regolith and delivers sediment to the channel (Simon and Guzman-Rios, 1990; Larsen *et al.*, 1999). Other hillslope weathering processes such as sheetwash, soil creep, and treefall-induced mass movement are prevalent but less important to the total sediment yield (Larsen, 1997). There is an abundance of clay in the deeply weathered soil and thick saprolite that is derived from both volcanoclastic and granodiorite bedrock (Frizano *et al.*, 2002; Schellekens *et al.*, 2004). However, there is typically little fine sediment that persists in the streams channels (Simon and Guzman-Rios, 1990). Flood discharges quickly wash fine sediment from the channel, and the streams are generally clear within a day of a large storm.

Landslides triggered by intense rains are common at upper elevations, particularly on areas underlain by granodiorite, and on hillslopes that exceed 12° gradient (Larsen and Torres-Sánchez, 1998). Landslides are capable of delivering very large boulders to the stream channels, and the corresponding volume of material transported is substantial (700 Mg km⁻² yr⁻¹ on granodiorite; 480 Mg km⁻² yr⁻¹ on volcanoclastics) (Larsen, 1997). The large majority (80–90%) of total sediment delivery to the streams is attributed to landslides, and the associated pulse of sediment delivery can locally alter the channel morphology. Since 1979, there have been numerous sliding events, and two large landslides have temporarily dammed permanent streams, persisting for a few weeks before being removed by stormflow (personal observation by F.N. Scatena).

The first-order drainage network consists of a dense, dendritic network of small ephemeral channels that range from leaf-filled swales to mossy cobble-lined channels that become active only during large rainfall events (Scatena, 1989; Schellekens *et al.*, 2004). Larger first-order perennial streams have channels dominated by boulders and clay and soil-lined channel banks. Second- and third-order streams have high-gradient reaches, exposed bedrock channels, matrices of large boulders interspersed with finer sediment, and periodic waterfalls (up to 30 m in height). Many of the upland streams are characterized by cascade and step-pool morphologies, whereas the lower reaches are plane bed and pool-riffle sequences (*sensu* Grant *et al.*, 1990; Montgomery and Buffington, 1997; Trainor and Church, 2003). Structural control of channel pattern is apparent in many places as witnessed by rectangular stream bends at fault intersections, streams following bedrock joints, and knickpoints at lithologic boundaries. Due to rapid decomposition, these channels lack the large coarse woody debris dams that influence the morphology of many channels in humid temperate environments (Covich and Crowl, 1990).

Fourth- and fifth-order streams occur only at lower elevations, flowing across the coastal plain as relatively gentle gradient pool-riffle sequences. Alluvial inset deposits, high-flow channels, floodplains, and terraces are common features in these lower reaches (Ahmad *et al.*, 1993, Clark and Wilcock, 2000). These larger lowland streams are gently meandering, are not constricted by bedrock, and have laterally migrating high-flow channels that indicate that the alluvial channels adjust in response to varying discharge and sediment supply.

Methods

A total of 238 stream cross-sections were surveyed in the summers of 2003–2006; 11 in the Río Blanco, 88 in the Río Espiritu Santo, 31 in the Río Fajardo, 91 in the Río Mameyes, and 17 in the Río Sabana (Figure 1). Cross-section locations were chosen to capture the entire range of elevation, drainage areas, and substrate type. The surveyed cross-sections are located on 1st to 5th order streams, have drainage areas between 0.1 km² and 79 km², and are located on average approximately every 30 m in elevation and 500 m in distance along the channel.

Cross-sections were surveyed at a straight uniform section within a reach. Relative distance and elevation were measured at evenly spaced intervals along a transect spanning from vegetated bank to bank, using a Sokkia Total Station Laser Theodolite (Set 530) (Sokkia, North America). In alluvial channels, cross-sectional geometry was measured at the bankfull stage, corresponding to the effective discharge of sediment (Leopold and Maddock, 1953; Wolman and Miller, 1960). However, the absence of floodplains in the mountainous reaches confounded the identification of bankfull stage. In steepland streams, cross-sections extended to the boundary of the active channel marked by the edge of perennial woody vegetation (shrubs and trees) and incipient soil development. A previous analysis of flow-frequency at gaged stream reaches indicates that these vegetation and soil indicators demarcate an active-channel boundary coinciding with a flood discharge that occurs at the same frequency of both the bankfull and effective discharge in adjacent alluvial channels (Pike and Scatena, 2010, in press). Using these riparian features as markers, active channel width, average depth, and cross-sectional area were calculated for each cross-section. Local channel slope was measured as the difference in elevation of the water surface over 10 uniformly spaced points

spanning approximately five channel widths upstream of the cross-section.

Active-channel (bankfull) discharge at each cross-section was estimated by a regional equation based on long-term stream gage data (Pike, 2006). At nine stream gages (with at least 10 years of record) in the study watersheds, the active channel discharge (as marked by the first occurrence of woody shrubs, trees, and soil) corresponds to a flood that is exceeded 0.16% of the time (0.6 days yr⁻¹), a recurrence interval of 90 days, and has an average unit discharge of 2.2 m³ s⁻¹ km⁻² (Pike and Scatena, 2010, in press). In this region, rainfall and runoff increase with elevation due to the precipitation gradient; higher elevation basins have more runoff than low elevation basins of comparable size. Thus, active channel discharge was best estimated as a function of drainage area multiplied by a linear model relating runoff to average basin elevation (Pike, 2006):

$$Q_{AC} = DA(0.0042 * E_{avg} + 0.406) \quad (1)$$

where Q_{AC} is the active channel discharge (m³ s⁻¹), DA is drainage area (km²), and E_{avg} is the average upstream elevation (m). Both of these variables are estimated for each reach using a 10 m digital elevation model (DEM) derived from a 10 m USGS contour map and a GIS-based flow accumulation algorithm.

Longitudinal profiles were constructed from original contours, as recommended by Wobus *et al.* (2006). Elevation, drainage area, slope, and active-channel discharge (Equation (1)) were estimated along the mainstem profile at a series of points intersecting the original 10 m contour lines; that is, spaced at every 10 m drop in elevation. Concavity was estimated similarly along the mainstem based on a best-fit power relationship between slope and drainage area:

$$S = kDA^\theta \quad (2)$$

where θ is the concavity index, k is a steepness coefficient, DA is drainage area (km²), and S is slope (m m⁻¹).

Downstream hydraulic geometry relationships were calculated by least-squares log-linear regressions between active channel discharge and channel geometry measurements. Active channel discharge (Q) correlates with active channel width (w), average flow depth (d), and mean velocity (v), such that: $w = c_1 Q^{b_1}$, $d = c_2 Q^{f_2}$, $v = c_3 Q^{m_3}$ (Leopold and Maddock, 1953). By conservation of mass, the product of the coefficients (c_1 , c_2 , c_3) and sum of the exponents (b , f , m) must equal 1; $c_1 \cdot c_2 \cdot c_3 = 1$, $b + f + m = 1$. Mean velocity was calculated as the active channel discharge divided by the channel cross-sectional area (Q_{AC}/A). Since this indirect calculation of velocity is a function of discharge, the strength of the regression equation was estimated as the correlation between the logarithms of active channel discharge and cross-sectional area.

Grain size in the active channel was estimated using a modified Wolman Pebble Count method (Wolman, 1954). Approximately 100 clasts were selected randomly by pacing across the width of the stream. The median diameter of each clast was measured, and classified into the following seven size categories: bedrock (no size), megaboulder (>2000 mm), boulder (256–2000 mm), cobble (64–256 mm), gravel (2–64 mm), sand (0.063–1 mm), and fines (silt/clay, 0.001–0.063 mm). From these grain size measurements, we determined the median grain size (d_{50}), coarse grain size (d_{84}), and the percent of bedrock exposed within the active channel.

Sediment mobility was calculated using the survey data and estimates of shear stresses. Sediment is considered mobile

when the dimensionless boundary shear stress exceeds the dimensionless critical shear stress. Boundary shear stress, τ (N m⁻¹), was calculated as:

$$\tau = \rho g R S \quad (3)$$

where ρ is water density (1000 kg m⁻³), g is acceleration due to gravity (9.8 m s⁻²), R is hydraulic radius (m), S is slope (m m⁻¹).

Dimensionless shear stress, τ^* , or Shields stress, required to mobilize the coarse sediment was estimated as:

$$\tau^* = \frac{\tau}{(\rho_s - \rho) g d_{50}} \quad (4)$$

where ρ_s is sediment density (2650 kg m⁻³), d_{50} is the median grain size (in m).

Many studies simply assume that river beds are mobile when Shields stress is greater than 0.03–0.07 (Buffington and Montgomery, 1997). However, recent analyses of data from laboratory flumes and natural streams show that the critical Shields stress of initial sediment motion increases with channel slope, indicating that particles of the same size are more stable on steeper slopes (Mueller *et al.*, 2005; Lamb *et al.*, 2008). This slope dependence is critical in this study because many of the reaches have slopes greater than 0.01 (when slope dependence is most significant). Dimensionless critical shear stress, τ^*_{cr} , was estimated from an equation developed for steep gravel-bed rivers by Mueller *et al.* (2005), that relates the dimensionless critical shear stress to slope (S):

$$\tau^*_{cr} = 2.18S + 0.021 \quad (5)$$

Equation (5) accounts for the variability in the critical dimensionless shear stress that is present because of excess bed roughness in steep reaches with high grain emergence, changes in local flow velocity, and turbulent fluctuations, as supported from data from numerous steep gravel-bed rivers (Lamb *et al.*, 2008).

Total stream power per unit channel length, Ω (W m⁻¹), is defined as:

$$\Omega = \rho g Q S \quad (6)$$

Stream power was calculated for each reach using the survey data and estimated active-channel discharge.

Results

Longitudinal profiles

The longitudinal profiles of each of the five rivers have unique shapes that are related to their underlying geology. A theoretical graded concave-upward profile is steepest in the headwaters, and displays a systematic downstream decline in slope. However, the profiles here are segmented by a series of convex protrusions and slope breaks that deviate from a systematic grade. For example, the volcanoclastic headwaters of the Río Fajardo and the contact metamorphic upper reaches of the Río Mameyes display a traditionally concave shape (Figure 2). Similarly, the alluvial reaches are well-graded and have few slope breaks. However, local factors also shape the profile. For example, the steep streams on volcanoclastic rocks have knickpoints that generally correspond to bedrock faults identified on USGS 1:20,000 geologic maps (Seiders, 1971a;

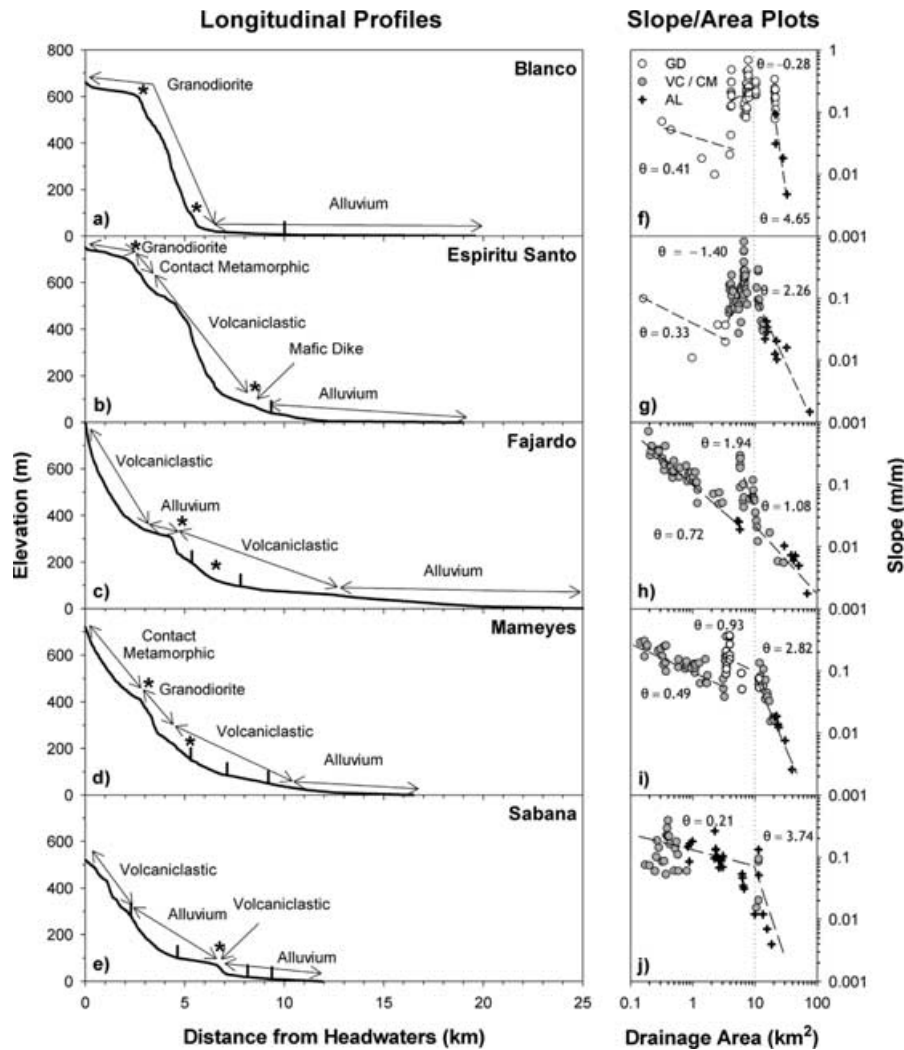


Figure 2. Longitudinal profiles of the main stem of each river highlighting the relationship between local profile shape and lithology (a–e). Mapped faults (vertical bars) are indicated. Slope–area plots (f–j), using points spaced every 10 m in elevation along the main stem, are shown to indicate changes in the concavity index (θ). Points are plotted according to the underlying lithology (white circle = granodiorite, grey circle = volcaniclastics and contact metamorphics, black cross = alluvium). The break where drainage area exceeds 10 km² is indicated by the dashed line. Concavity values are derived from best-fit lines based on the lithology in headwater reaches, and for points where $DA > 10$ km². Black asterisks in the longitudinal profile plots indicate changes in the slope–area relationships.

Briggs and Anguilar-Cortez, 1980). Also, small convexities in the longitudinal profiles correspond to locally exposed outcrops. Most striking are the anomalously convex profiles of granodiorite streams. The headwaters of granodioritic Río Blanco are unusually flat before cascading steeply down the side of the batholith and leveling out along the alluvial coastal plain (Figure 2). The inflection point where the stream sharply steepens occurs at the edge of a non-glacial hanging valley. This flat form is also seen in the headwaters of the Río Espíritu Santo, where it is also associated with granodiorite bedrock.

Where the stream flows across one lithology, the longitudinal profile is locally graded and has a traditional concave shape. Where the stream flows over two or more rock types, there is often a slope break at the contact, especially where the adjoining rocks have varying resistance to erosion. The boundary between contact metamorphic rocks and other lithologies, as on the mainstem of the Río Mameyes and Río Espíritu Santo, is accompanied by a pronounced convexity (Figure 2). Furthermore, slight changes in the composition of the volcaniclastic rocks, from a sandstone unit to a mudstone unit, are often the site of waterfalls and/or steep gradients (personal observation). Similar notable breaks occur where

upland alluvial formations, typically terrace deposits, merge with volcaniclastics, such as a knickpoint on the Río Fajardo that occurs at the boundary between a mid-elevation structural bench and the surrounding volcaniclastics. Also, the lowest elevation waterfall in the region occurs at the transition where the Río Sabana flows across a locally exposed volcaniclastic formation approximately 7 km from its headwaters.

The concavity index (θ) of the mainstem of each river profile, calculated from slope–area relationships, is related to both the underlying rock type and drainage area (Figure 2f–2j). Concavity values for headwater portions of the profiles ($DA < 10$ km²) are starkly different between areas underlain by granodiorite and those by volcaniclastics. Concavities of headwater profiles that are primarily on volcaniclastic and contact metamorphic rocks (Río Fajardo, Río Mameyes, and Río Sabana) are in the low to moderate range ($\theta = 0.15$ to 0.72), with slightly higher concavities after significant lithological breaks on the Río Fajardo ($\theta = 1.94$) and Río Mameyes ($\theta = 0.93$) profiles. In contrast, channel profiles that having significant headwater portions on granodiorite (Río Blanco and Río Espíritu Santo), concavity values are moderate ($\theta = 0.33$ to 0.41) along the gently-sloped reaches and negative/

convex ($\theta = -0.28$ to -1.40) where the channel steepens. There is a break in the slope-area relationship at approximately 10 km^2 along each profile. Consequently, lowland and alluvial portions of the profiles ($DA > 10 \text{ km}^2$), are all strongly concave ($\theta = 1.08$ to 4.65), regardless of the underlying lithology.

Hydraulic geometry

Downstream hydraulic geometry relationships for the active channel width, hydraulic radius, and mean velocity were calculated using the cross-sectional data and estimated active channel discharge (Figure 3). The coefficient of determination (r^2) for these hydraulic geometry relationships are 0.71 for width, 0.21 for depth, and 0.66 for velocity (as determined by the relationship with cross-sectional area) (Table 1). Discharge displays a strong power-law relation with both width and velocity, but not depth. The active channel systematically widens in the downstream direction, despite potential constriction from bedrock outcrops and confined valley walls. Yet the streams do not deepen substantially downstream. Instead, they display strong local variation. Comparably deep pools and shallow riffles are observed in both headwater and lowland reaches.

Downstream hydraulic geometry exponents for all basins are 0.33 for width, 0.12 for depth, and 0.55 for velocity. With increasing discharge, width increases at approximately three times the rate of depth. This implies that the width/depth ratio similarly increases in the downstream direction, and that the channel form changes from a triangular 'v'-shape (low w/d ratio) within the headwaters to a more rectangular (high w/d ratio) form near the mouth.

DHG relationships for individual watersheds show general consistency among basins (Table I). The width exponents range from 0.24 to 0.37, the depth exponents from 0.02 to 0.17, and the velocity exponents from 0.52 to 0.62. The differences in coefficients and exponents are not strongly correlated to basin-scale factors such as catchment size or geology. For each watershed, the r^2 values for width and velocity relationships are >0.5 , but less than 0.5 for depth relationships. Consequently, DHG is considered well-developed for all the watersheds.

For comparison, average DHG exponents for many alluvial rivers worldwide are 0.5 for width, 0.4 for depth, and 0.1 for velocity (Park, 1977). DHG exponents in mountain streams deviate slightly from the world average by having a lower width exponent and greater velocity exponent, with average values of 0.36 for width, 0.38 for depth, and 0.20 for velocity (Wohl, 2004). The Luquillo streams have a width exponent comparable with other mountain streams, but both the lowest known depth exponent and the highest velocity exponent for a mountain stream.

Grain size

The grain size of bed material varies widely throughout the watersheds, but is clearly related to the underlying rock type (Figure 4). For example, long stretches of step-pool sequences composed of boulders up to several meters in diameter are present in a steep upland tributary of the Río Espíritu Santo underlain by volcanoclastic rock. In contrast, the headwaters of the Río Blanco are composed mostly of mobile sand that is weathered from granodiorite interspersed with large boulders. A typical lowland reach is composed of cobbles and gravels derived both from the surrounding alluvium and transported

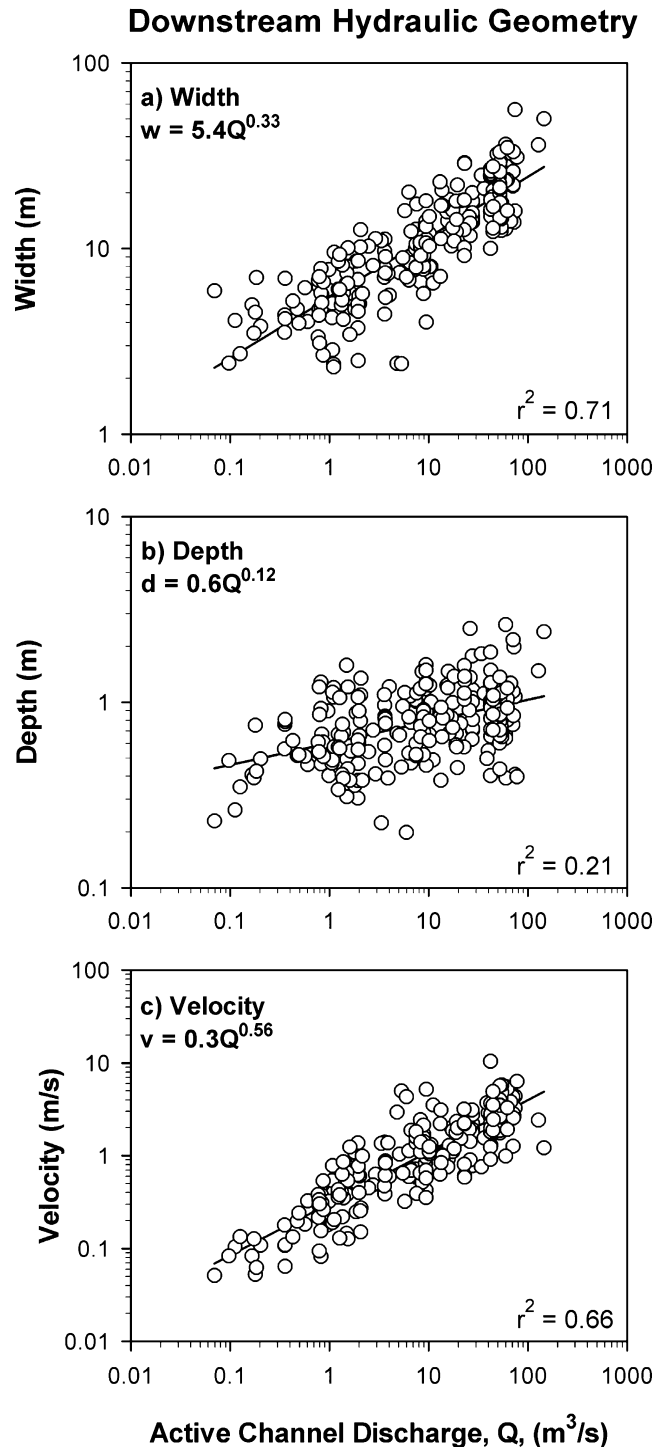


Figure 3. Downstream hydraulic geometry relationships between active channel discharge and width (a), depth (b), and velocity (c) using data from all surveyed reaches.

from upper reaches. The largest clasts observed in the river channels are slabs of volcanoclastic rock and granodiorite corestones that reach 15 m in diameter. These are so large and immobile that they are hydraulically indistinguishable from bedrock.

Unique distributions of grain sizes are observed on different lithologic types. Using the pebble-count data from each study reach, all measured grains (excluding bedrock) were sorted into logarithmically distributed bins (2ϕ intervals) for each major mapped lithology at the measured station (Figure 5). Streambed material on volcanoclastic rocks has a high frequency of cobble and boulder sized-sediment (64–1028 mm),

Table 1. Downstream hydraulic geometry divided by watershed. Coefficients, exponents, and coefficient of determination (r^2) are between the active channel discharge and each corresponding channel geometry variable

Watershed	width			depth			velocity			
	c_1	b	r^2	c_2	f	r^2	c_3	m	$*r^2$	n
Blanco	4.0	0.37	0.65	0.6	0.09	0.05	0.4	0.54	0.54	11
Espiritu Santo	5.9	0.30	0.63	0.6	0.13	0.22	0.3	0.57	0.55	88
Fajardo	4.9	0.35	0.73	0.9	0.02	0.02	0.2	0.62	0.73	31
Mameyes	5.2	0.35	0.80	0.6	0.13	0.33	0.3	0.52	0.77	91
Sabana	7.7	0.24	0.50	0.4	0.17	0.30	0.3	0.58	0.57	17
ALL	5.4	0.33	0.71	0.6	0.12	0.21	0.3	0.56	0.66	238

* r^2 for velocity relationships calculated from discharge versus cross-sectional area

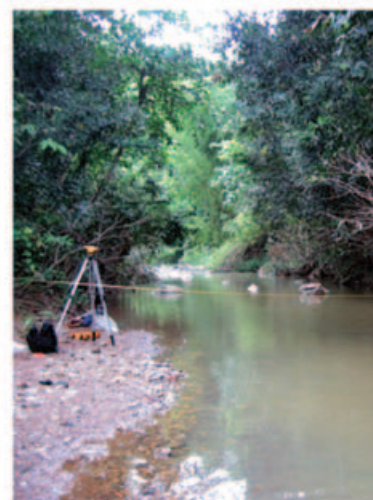
a) Volcaniclastic - Upland
Rio Espiritu Santo



b) Granodiorite - Upland
Rio Blanco



c) Alluvium - Upland
Rio Sabana



d) Volcaniclastic - Downstream
Rio Mameyes



e) Granodiorite - Downstream
Rio Blanco



f) Alluvium - Downstream
Rio Fajardo



Figure 4. Upstream views of typical reaches throughout the basins. The grain size varies with both the lithology and the position along the stream profile. The average channel width / median grain size, d_{50} , for each reach are: (a) 13.2 m/480 mm; (b) 6.8 m/1 mm; (c) 11.0 m/60 mm; (d) 32.3 m/150 mm; (e) 14.0 m/330 mm; (f) 25.4 m/70 mm. This figure is available in colour online at www.interscience.wiley.com/journal/espl

but also contains lesser proportions of large boulders (>1028 mm) and gravel. Field observations suggest that different volcaniclastic formations have varying proportions of large boulders that are dependent on the formation thickness. Contact metamorphic rocks display a distribution with fewer large boulders and more sand than their unmetamorphosed equivalents. Mafic dikes have a high proportion of large boulders,

which is unique given the lower elevation. Granodiorite streams have a bimodal grain-size distribution composed primarily of sand and large boulders. Alluvial streams contain an abundance of cobble and gravel-sized grains.

Megaboulders (boulders >2000 mm diameter) are a relatively common feature in the channels. Presumably, the megaboulders are corestones that are weathered directly from

bedrock along fracture planes, and subsequently deposited into the stream channels by landslides. The abundance of megaboulders in the channel correlates with the slope of the adjacent hillsides. Where the adjacent hillslope exceeds a threshold of 12° , there is potential for landslides and possible deposition of megaboulders into the stream channel (Larsen and Torres-Sánchez, 1998). Conversely, below this hillslope threshold, the hillside is relatively stable. Among 45 reaches having shallow adjacent hillslopes that do not exceed 12° , no megaboulders were observed (Figure 6). Of the remaining 193 reaches adjacent to steep hillsides exceeding the 12° threshold, about half (89 reaches, 46%) have megaboulders present. Thus, megaboulders are only found in about half of these reaches that are considered landslide-prone areas based on slope and lithology. However, not all reaches in landslide-prone areas have megaboulders.

Most of the reaches have a relatively stable framework of large boulders that do not appear to move, as evidenced by a thick moss-covering and occasional tree growth on the boulders' surfaces. Yet within this matrix, there is an abundance of smaller loosely packed gravel and sands that is transported during floods (personal observation). Average dimensionless shear stress in these channels varies considerably but generally decreases as a function of active-channel channel discharge (Figure 7a). Critical dimensionless shear stress, as estimated by Equation (5), also decreases with increasing active-channel discharge (Figure 7b). This suggests that many of the headwater reaches have additional flow resistance which leads to an increased threshold to initial sediment motion. Excess dimensionless shear stress (τ^*/τ_c^*) does not vary systematically with active-channel discharge (Figure 7c). The average dimensionless boundary shear stress at the active channel discharge exceeds the dimensionless critical shear stress to mobilize the d_{50} in approximately 45% of the study reaches (Figure 7c). That is, the median sediment size in nearly half of the surveyed reaches can potentially be mobilized during the active channel flood that occurs several times per year. The lack of scaling of excess shear stress with discharge also suggests that at the active channel discharge, the channels are generally at the sediment transport threshold – similar to many alluvial channels.

Given the variety of grain sizes and range of rock types, the pattern of grain size throughout the basin is not immediately apparent. The median diameter (d_{50}) of particle size at a reach correlates poorly with drainage area ($r^2 = 0.002$, $P = 0.83$), and significantly but only moderately with either slope ($r^2 = 0.20$, $P < 0.01$) or stream power ($r^2 = 0.30$, $P < 0.01$). However, a log-linear plot between the ratio of d_{50} to drainage area ($d_{50}:DA$) and slope (*sensu* Hack, 1957) yields a stronger and highly significant correlation ($r^2 = 0.74$, $P < 0.001$) (Figure 8). A similar significant relationship was found for the d_{84} data ($r^2 = 0.72$, $P < 0.01$).

$$S = 0.007 \left(\frac{d_{50}}{DA} \right)^{0.55} \quad \text{or} \quad d_{50} = DA \left(\frac{S}{0.007} \right)^{18} \quad (7)$$

$$S = 0.0026 \left(\frac{d_{84}}{DA} \right)^{0.57} \quad \text{or} \quad d_{84} = DA \left(\frac{S}{0.0026} \right)^{175} \quad (8)$$

These two relationships state that grain size is a function of both drainage area and slope. At a given drainage area, grain size is proportional to 1.8 power of slope. Conversely, if slope remains constant, then grain size is directly proportional to drainage area.

Equation (7) was used to estimate median grain size along the profile of the river, using drainage area and slope derived from the 10 m DEM (Figure 9). The predicted grain size

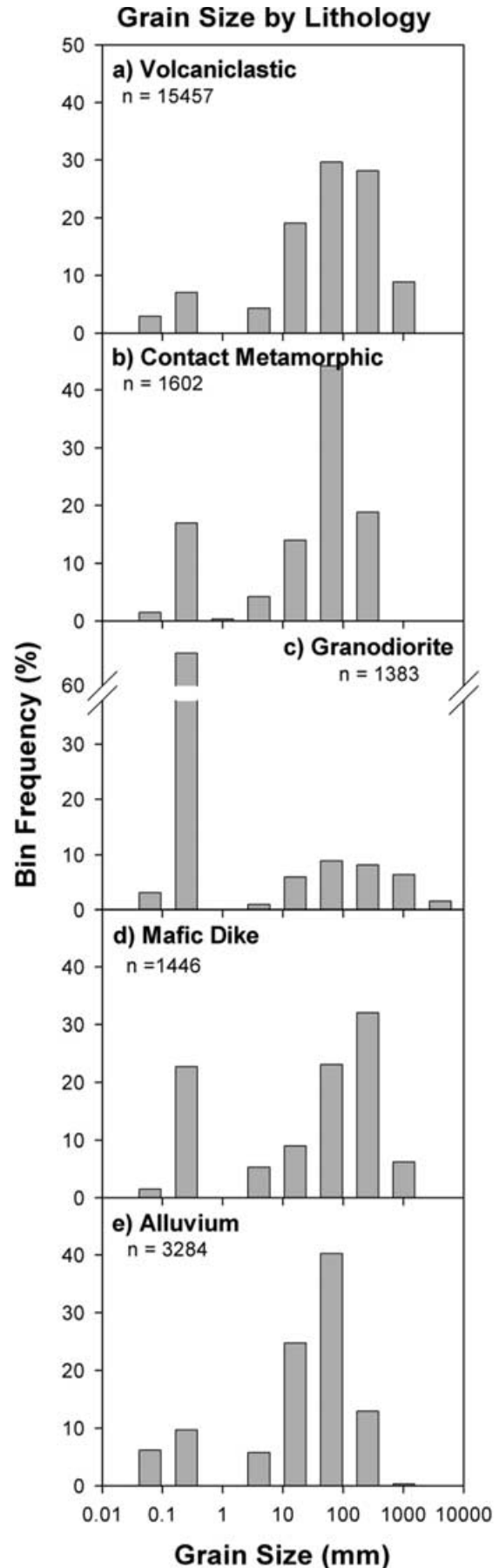


Figure 5. Grain size histograms for all measured clasts at all locations, grouped by lithology. Bins are logarithmically distributed at 2ϕ intervals.

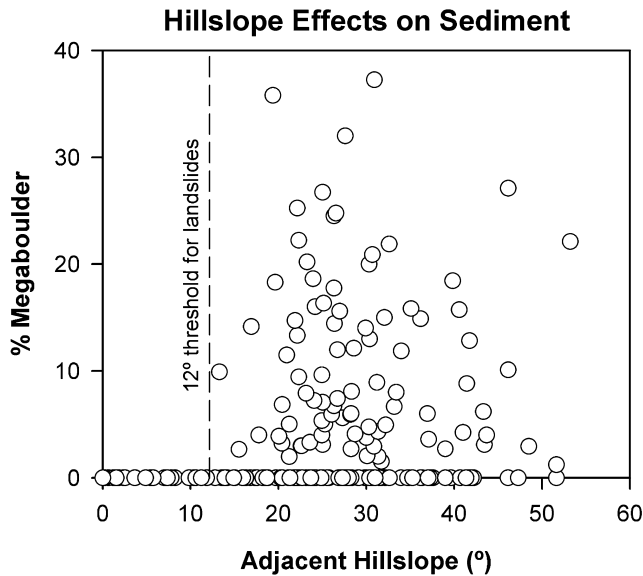


Figure 6. The steepness of the adjacent hillslope vs. the percentage of megaboulders (boulders >2000 mm in the channel). Megaboulders are present when the hillslope exceeds 12°. This is also the slope threshold for landslides—the process that presumably delivers these large boulders to the channel.

function is jaggedly shaped due to variations in slope. Yet the resulting pattern shows a distinct downstream coarsening from the headwaters to a mid-basin maximum. This is followed by subsequent downstream fining. Typically, cobble and small boulder-sized sediment in the headwaters are replaced by very large boulders (>2000 mm) in the mid- to upper-elevation steepland channels. Along the mainstem, this maximum typically occurs at approximately 5 km from the headwaters, or approximately a quarter to a third of the profile length. As the slope declines towards the lowland reaches, the sediment gradually fines as well.

The measured grain size data for reaches along the mainstem show general agreement with the predicted grain size (Figure 9). The data best display the coarsening to fining trend in those rivers where surveyed reaches span a range of slopes and drainage areas (Río Espíritu Santo, and Río Sabana). On other rivers, such as the Río Fajardo, the measured grain sizes do not vary as widely because the surveyed reaches had comparable slopes.

The largest discrepancy between the actual grain size and the predicted grain size occurs in the steepest reaches (Figure 9). Here, the grain size function predicts boulders in excess of 10 m. These reaches are generally waterfalls and cascades where the largest sediment is practically indistinguishable from bedrock, or is bedrock itself, and consequently cannot be quantified by a measurable diameter. Field observations did indicate notably larger grains in the mid-basin steep reaches. Data from waterfalls and cascades on tributaries also show that the largest grains are found in the steepest reaches with moderate drainage areas.

Stream power

Total stream power along the mainstem displays a peaked pattern in the downstream direction (Figure 9). Given that total stream power is the product of discharge and slope, stream power is low in the headwaters where slopes are steep but there is minimal discharge. It is similarly low near the mouth where there is high discharge but gentle slopes. Thus, stream

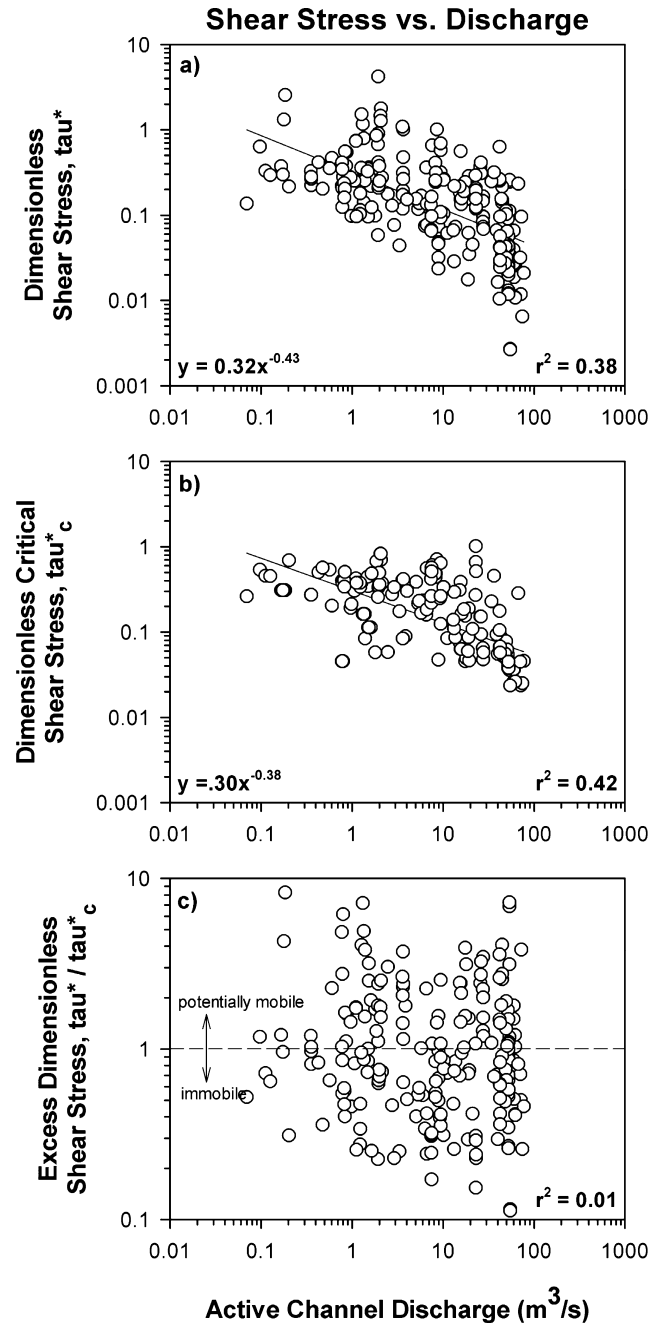


Figure 7. Relationship between active channel discharge and dimensionless shear stresses and critical shear stresses. (a) Dimensionless boundary shear stress is negatively correlated with active channel discharge, rather than constant. (b) Dimensionless critical shear stress (a function of slope) also decreases with active channel discharge. (c) Excess shear stress (as the ratio of dimensionless shear stress to critical shear stress) does not vary systematically downstream with discharge, suggesting that the channels waver around the threshold for sediment transport. Approximately 45% of the reaches have presumably mobile substrate ($\tau^* > \tau_c^*$) at the active channel discharge.

power peaks at an intermediate distance where a combination of adequate discharge and steep slopes generate maximum power. In these streams, the location of this stream power peak typically occurs at mid- to upper-elevations and at a downstream distance approximately a quarter to a third of the profile length, also corresponding to the maximum slope and maximum grain size. The magnitude of the peak varies according to the watershed, with the larger and steepest watersheds (such as the Río Blanco) having greater maximum stream power.

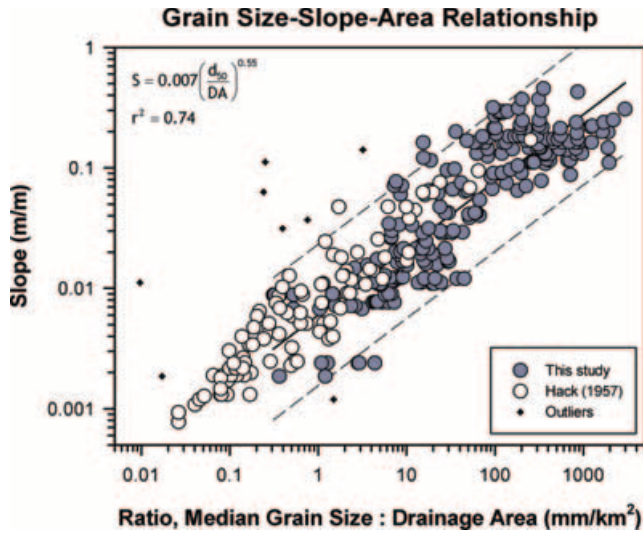


Figure 8. The highly significant relationship ($r^2 = 0.74$, $P < 0.0001$) between median grain size (d_{50}), drainage area, and slope. Data from this study (dark circles) are shown alongside data representing humid streams in Maryland and Virginia from Hack (1957, white circles) where this relationship was first published. The solid line is a least-squares regression used as the basis for estimating grain size for a given drainage area and slope (Equation (4-7)), and dotted lines represent the 95% prediction interval. Outliers from this study (crosses) were removed from regression on the basis of uncertainty of either average grain size estimation or slope measurement. This figure is available in colour online at www.interscience.wiley.com/journal/espl

The ratio of total stream power to the coarse grain size (Ω/d_{84} , $\text{W m}^{-1} \text{m}^{-1}$) shows the relative influence of hydraulic forces (as stream power) to lithologic resistance (coarse grain size). Along the mainstems, this ratio generally shows a positive trend in the downstream direction (Figure 9), indicating the relative dominance of stream power in the lowland reaches, and strong resistance by coarse grains in the headwater reaches. Using a threshold of $10\,000 \text{ W m}^{-1} \text{m}^{-1}$ to differentiate between supposed alluvial conditions and lithologic controls (Wohl, 2004), it is apparent that the transition from strong lithologic control to more alluvial conditions occurs approximately a third to a halfway down the length of the mainstem. Furthermore, those watersheds having granodiorite substrate in the headwaters (Río Blanco and Río Espíritu Santo) have a Ω/d_{84} in excess of this threshold, indicating they have alluvial conditions in their sandy headwater reaches. The longitudinal peaks and transitions in grain size, stream power, and the ratio of total stream power to the coarse grain size also corresponds to the location where the drainage area exceeds 10 km^2 , which marks a break in the slope–area relationship (as shown in Figure 2).

Discussion

The results presented above indicate that the streams of the Luquillo Mountains have an intricate connection between the underlying lithologic and hydraulic controls, and the resulting

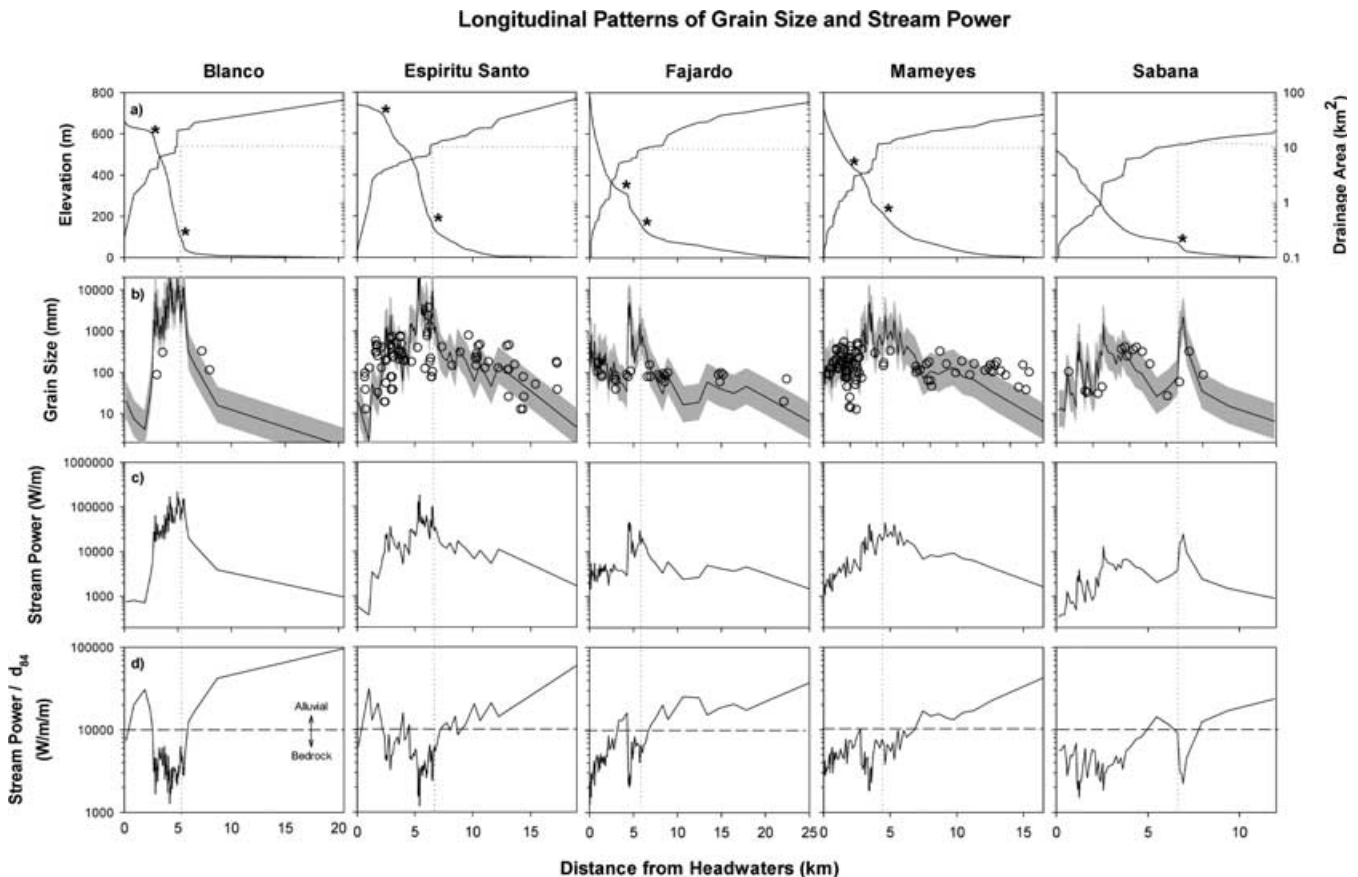


Figure 9. Downstream changes in elevation and drainage area (a), median grain size (b), stream power (c), and the ratio of stream power to coarse grain size (d) along the main stem of each river. For trends in grain size, 95% prediction intervals are shown in grey, and measured data are plotted as circles. Both median grain size and stream power display a peaked pattern in the downstream, the location of which coincides with steep reaches having sizeable flow. The ratio of stream power to coarse grain size increases in the downstream direction, suggesting the increased relative influence of hydraulic forces over lithologic controls downstream. A threshold of $10\,000 \text{ W m}^{-1} \text{m}^{-1}$ (dashed line) differentiates alluvial conditions from bedrock controls. The location where drainage area exceeds 10 km^2 is indicated by the dotted line, and black asterisks indicate significant changes in concavity (from Figure 2).

profile shape, grain size distribution, channel geometry, and channel energetics. Local-scale geologic factors such as the rock type, exposed in-channel outcrops, and bedrock faults are seemingly dominant in determining the shape of the longitudinal profiles. Different lithologies correspond to local variations in profile slope and concavity. They also weather into unique particle sizes, and are associated with specific channel geometries. Large immobile boulders are deposited in the channel by landslides. Yet increasing discharge and hydraulic controls on sediment transport override these lithologic influences and give rise to basin-scale patterns. This is evidenced in that channel geometry and grain size are also strongly related to slope and discharge (Figures 3, 8 and 9). Here we discuss the importance of each of these basin-scale patterns and the implications for the dynamics of tropical mountain streams.

The longitudinal profile of each of the study rivers, although generally concave-upward, display fragmented patterns consistent with lithologic control. The rivers have slope breaks, knickpoints, and profile convexities that correlate with different rock types and structural features of the underlying bedrock. The relative strength of different bedrock types during fluvial incision can yield such segmented profiles (Brocard and van der Beek, 2006). Chemical and physical weathering, as well as debris flows, are the dominant processes of bedrock incision in these rivers, as noted in other mountainous drainages (Stock and Dietrich, 2006). Many of the common processes of river incision into bedrock, notably plucking, macro-abrasion, wear, and cavitation (Whipple, 2004) are rarely observed.

The concavity of the longitudinal profiles is related to underlying bedrock and hillslope processes. Whipple (2004) discusses potential controls on channel concavity. Low to moderate concavities (<0.7), such as that are seen in many of the headwaters of those rivers draining volcaniclastic, are associated with short, steep drainages importantly influenced by debris flows. Convex profiles (negative concavity), seen along primarily granodiorite streams, are typically associated with abrupt knickpoints owing either to pronounced along-stream changes in substrate properties (VanLaningham *et al.*, 2003) or to spatial or temporal differences in rock uplift rate (Whipple, 2004). Extreme concavities (>1.0), present along the lowland and alluvial reaches, are associated with transitions from incisional to depositional conditions. Thus, the transition from low concavity to high-concavity at approximately 10 km^2 marks a transition from dominant colluvial and hillslope processes and incisional channels to depositional alluvial-type channels.

Brummer and Montgomery (2003) noted a similar break in the slope–area relationships at a drainage area of 10 km^2 in some coastal temperate streams, contending that the associated change in concavity reflected a shift from dominant colluvial processes in headwater channels to alluvial processes in lowland channels. In fact, many key transitions in channel pattern occur around a drainage area of approximately 10 km^2 – the supposed threshold separating the dominance of colluvial and alluvial forces. The peaks in the longitudinal patterns in grain size and stream power, as well as the transitions in the relative strength of fluvial forces over lithologic resistance occur around this point (Figure 9). This suggests that although there are local variations in headwater channel patterns owing to transitions in lithology, the most drastic channel changes at the basin scale occur where the streams leave the mountains and begin to enter the coastal plain.

Downstream hydraulic geometry is considered well developed in all of these basins, despite the influence of non-fluvial processes, differences in lithology, and local structural features. Mountain rivers are considered to have well-developed

DHG when the coefficient of determination (r^2) between discharge and at least two of the three hydraulic variables is 0.5 or greater (Wohl, 2004). The high r^2 values for width (0.71) and velocity (0.66) relationships in the Luquillo streams satisfy this criteria, so that the DHG for the stream network is considered well-developed.

The well-developed hydraulic geometry can be attributed to the strong influence of fluvial forces over lithologic resistance, as reflected in the ratio of stream power to grain size. Wohl (2004) found that mountain streams have well developed DHG when the ratio of total stream power to the coarse grain size, Ω/d_{84} , is greater than $10\,000 \text{ W m}^{-1} \text{ m}^{-1}$. Above this threshold, the river presumably has enough power to rework the coarse sediment and adjust channel parameters in response to downstream changes in discharge. Below this threshold, combinations of low stream power or large grain sizes inhibit the river from developing strong DHG relationships. The average Ω/d_{84} ratio of all surveyed reaches in the streams of the Luquillo Mountains is approximately $14\,000 \text{ W m}^{-1} \text{ m}^{-1}$, or slightly above the threshold. Although the average grain size is very large, so is the average stream power. Thus, the combination of high discharge and steep slopes generates sufficient stream power to overcome lithologic controls and adjust channel geometry accordingly over time.

Clark and Wilcock (2000) noted a DHG reversal in the lowland alluvial reaches in some of these streams. Values of channel width, depth, and velocity either decreased or were constant in the downstream direction in the lowland reaches. This hydraulic geometry reversal trend was found on coastal plain alluvial reaches along the mainstem, or approximately the lower 33% of the mainstem profile length. The authors attribute this reversal to historic and modern land-use changes. Apparently, the shift from forest to agriculture to urbanization over 400 years altered the sediment supply and flow regime. Net aggradation of sediment during periods of land-clearance and recent net degradation from heightened runoff due to urbanization have altered the balance that maintains channel geometry. However, our data confirms that this reversal is strictly confined to the lowland alluvial reaches. Hydraulic geometry remains relatively well developed at the basin scale that spans four orders of magnitude in discharge.

The poor correlation between depth and discharge across all watersheds suggests that local factors are a strong determinant of channel form. Bedrock outcrops, scour pools, and the accumulation of large boulders can all locally determine depth. To compensate for the small increase in depth, these streams increase drastically in velocity with increasing downstream discharge. We document the largest downstream velocity exponent reported for a mountain stream. This sizeable downstream increase in velocity may be a result of the basin physiography and variations in flow resistance. These island streams generally have shorter and more truncated profiles than streams on continental land masses. Yet the short coastal plain is still relatively steep so that the flood waters flow rapidly to the ocean with minimal resistance. Despite having faster average flow, the downstream reaches are not the most energetic. Rather, upland streams that have lower average velocity, but greater slope, shear stress, and flow resistance, expend the greatest amount of energy.

The peaked pattern of grain size along the profiles of these rivers stands in contrast to a common systematic downstream fining trend in many alluvial rivers (Paola and Seal, 1995; Pizzuto, 1995). However, a similar systematic headwater coarsening pattern has been noted in several mountain basins in western Washington (Brummer and Montgomery, 2003). In both western Washington and the Luquillo Mountains, grain

size and stream power maxima occur at approximately the same location as the transition from debris-flow and landslide dominated channels to fluvially dominated channels. This suggests that a tendency for downstream coarsening may be ubiquitous in headwater reaches of mountain drainages where debris flow processes set the channel gradient. Apparently, when landslides dominate the transport and routing of sediment in low-order headwater channels, a coarsening trend occurs. Downstream fining occurs as fluvial forces override colluvial forces as the driving sediment transport process in high-order alluvial channels. These observations suggest that basin-wide trends in d_{50} are also in part hydraulically influenced by variations in stream power, as well as by landslide deposits.

There is a complex interaction between profile slope, grain size, drainage area and lithology, as noted by Hack (1957; 1960). Data from this study follow a similar relationship between these three variables (Figure 8) as data from temperate piedmont streams in Maryland and Virginia (Hack, 1957). The streams in Luquillo display the same adjustment between the grain size, drainage area, and slope as more gentle gradient streams in a very different physiographic region. The same basic relationship holds even though the Luquillo streams have steeper slopes and consequently greater d_{50} :DA ratios. Furthermore, rock type does not factor into this relationship, so that reaches on all lithologies display the same relationship among the three variables. The causal mechanisms associated with this relationship (i.e. whether slope is influenced by both the size of the sediment and discharge, or whether slope and discharge determine the grain size) is seemingly time-scale dependent (*sensu* Schumm and Lichty, 1965). Alluvial channels can adjust slope in response to transport capacity and sediment supply such that slope is a dependent variable related to water and sediment discharge, and grain size. Yet in the steep headwater bedrock channels where non-fluvial forces dominate, slope is generally imposed by lithology, and becomes an independent variable over the timescales of channel geometry adjustment. Channel sediment is shaped by persistent short-term fluvial and colluvial processes that organize the bed surface upon a slope set by longer-term erosion processes (Scatena, 1995).

Despite the abundance of large boulders throughout the basin, much of the interstitial bed-material among these boulders is readily mobile. The lack of correlation between excess dimensionless shear stress and discharge (Figure 7) suggests that throughout the stream network, these are 'threshold channels' that are capable of mobilizing moderately-sized sediment during bankfull floods. The constancy of excess shear stress is a feature commonly associated with alluvial channels that can readily adjust slope (Dade, 2000). However, Luquillo streams are evidently adjusted to be threshold channels, despite a geologically-imposed slope. Yet on longer time scales, the profile slope of these upland channels changes over the course of drainage network evolution. The upland channels adjust slope to the underlying lithology and consequently influence the type of sediment that is delivered to the channels. The combination of these processes and scales suggest that the resulting channel morphology is not exclusively controlled by a single factor.

Conclusion

The morphology of the stream channels in the Luquillo Mountains are influenced by a combination of both local lithologic controls and strong hydraulic forces. Slope and grain size in many headwater areas are imposed by properties

of the underlying lithology and coarse sediment delivery by landslides. Longitudinal profiles and concavity are strongly related to lithologic boundaries. At the reach scale, non-fluvial factors such as bedrock outcrops, knickpoints, and fault bends locally affect the channel morphology. Hillslopes are strongly linked to channel dynamics and colluvial processes are dominant in many headwater areas.

Within the framework set upon by local and non-fluvial constraints, there are many basin-scale patterns that indicate these streams function similar to some fully alluvial rivers. The presence of strongly developed hydraulic geometry relationships, grain size patterns organized to slope and discharge, and high stream power relative to channel resistance indicate the influence of overruling fluvial forces. Furthermore, excess dimensionless shear stress at bankfull wavers around the threshold for sediment mobility indicating the river is able to systematically transport sediment and organize its own morphology. These basin-scale patterns attest to the ability of the forceful flow regime generated by the humid tropical climate to sculpt mountainous streams that share some commonalities with alluvial rivers.

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