THE CONTROL OF GRAIN-SCALE MECHANICS ON CHANNEL FORM, LANDSCAPE DYNAMICS, AND CLIMATIC PERTURBATIONS IN GRAVEL-BEDDED RIVERS

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Dedicated to Jack and Roberta Phillips

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ABSTRACT

THE CONTROL OF GRAIN-SCALE MECHANICS ON CHANNEL FORM, LANDSCAPE DYNAMICS, AND CLIMATIC PERTURBATIONS IN GRAVEL-BEDDED RIVERS

Colin B. Phillips

Douglas J. Jerolmack

Landscapes evolve over millions of years, through the complex interplay of climate and tectonics. Mountains in particular represent a staggering range of spatial and temporal scales, challenging our ability to understand how the landscape is sculpted. Mountains do not simply disappear by bulk denudation. The key process of river incision results from the entrainment, displacement, and collision of coarse particles with the bed; a phenomenon known as bed load transport. This dissertation seeks to elucidate how bed load transport in natural rivers is driven by floods, to provide a mechanistic connection between climate and landscape evolution. Field surveys of coarse particle displacement and channel geometry are combined with hydrological time series, to study the interaction between floods and bed load dynamics, and their implications for channel form. Results from tagged cobbles demonstrate that mean particle displacement is proportional to applied fluid momentum in excess of the threshold of motion, while dispersion of tracers is superdiffusive due to the burial and excavation of cobbles. These field surveys reveal that particle motion remains in a state of partial transport for a diverse population of flows, and that particle sorting and transport distances closely match theory developed from small-scale laboratory experiments. Analysis of hydrological time series shows that the threshold of particle motion truncates the distribution of applied stress, resulting in thin-tailed distributions of forcing for flows above the threshold of motion. This analysis further shows that, because a coarse-grained river adjusts its geometry so that the flow at the banks is at the threshold of motion, the probability of experiencing larger stresses diminishes exponentially. Field surveys of channel geometry and particle size reveal that the geomorphological impacts of urbanization are reduced for coarse-grained channels adjusted to frequent sediment transport events. Taken together, these observations indicate that the threshold of particle motion represents a firstorder control on the influence of climate on river dynamics, and the landscapes through which they flow.

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CHAPTER 1 : Introduction

The mountainous landscapes of the Earth represent the complex interactions of tectonics, climate, and life. The dynamic interplay of climate and mountains has received considerable attention (e.g., Molnar and England, 1990; Bull, 1991; Ruddiman et al., 1997; Whipple et al., 1999; Hartshorn et al., 2002; Montgomery and Brandon, 2002), though we are just beginning to understand the myriad of ways in which life and landscapes interact (Reinhardt et al., 2010). Perhaps due to the staggering spatial and temporal scales at which mountains evolve, many approaches to deciphering the link between them and climate remain broadbrush and empirical (e.g., Anderson, 1994; Tucker and Slingerland, 1997; Peizhen et al., 2001). Over geologic time, mountain ranges are slowly worn away due to the persistence of rivers and glaciers. Though glaciers may be more effective at carving mountain ranges, eventually a river must transport the material to the ocean where it may be archived in the stratigraphic record (*Paola et al.*, 1992a). Thus, rivers represent one of the few threads through which climatic signals can propagate across the entire landscape. For example, rivers propagate the rise and fall of the oceans into mountain ranges, as pulses of incision and aggradation, respectively (Crosby and Whipple, 2006). Conversely, rivers represent the mechanism through which increased erosion brings sediment from the mountains to the oceans. The large timescales involved prevent the direct observation of long-term landscape evolution, however we can turn to numerical models – implementing approximations of known physics and empirically-derived transport laws – to speed up the process (Tucker and Hancock, 2010 and references therein). At large timescales, numerical models have explored the dynamics of generating stratigraphy, evolving statistically realistic landscapes, and assessing their response to climatic forcing (Paola et al., 1992a; Tucker and Slingerland, 1997; Tucker and Bras, 2000). However, not all signals and depositional records represent climatic external (allogenic) forcing, as rivers are not linear filters. Rivers may possess internal (autogenic) dynamics that shred any imprint of allogenic forcing (Kim and Jerolmack, 2008; Jerolmack and Paola, 2010; Finnegan and Dietrich, 2011; Jerolmack, 2011). Further complications arise through the power-law and potentially heavy-tailed scaling of floods (*Turcotte and Greene*, 1993), in that the distribution of floods is not necessarily well behaved. The non-linearity inherent in both climate and rivers presents a complicated picture to say the least. Thus, deciphering the interaction between external forcing and landscape evolution from the long-term record through stratigraphic observations and modeling remains a grand challenge (*Jerolmack and Paola*, 2010; *Lague*, 2014).

The long-term denudation of mountains is rate-limited by the sediment mass flux leaving the system, which occurs through the entrainment, displacement, and deposition of sediment particles. Through the study of coarse sediment transport – the process of bed load - this thesis attempts to elucidate a single component of the interaction between climate (external forcing) and mountain landscapes. In most cases, suspended sediment represents the majority of the mass leaving a landscape through rivers (Milliman and Syvitski, 1992; Willenbring et al., 2013); however, it is bed load that chips away bed rock through innumerable collisions, and thus sets the limiting rate of incision (Sklar and Dietrich, 2004). Bed load also controls the geometry of gravel-bedded rivers (*Parker*, 1978, 1979) (In this thesis, gravel is taken to be any grain larger than 10 mm). Bed load transport is not without its complexities; even under steady flow, it is known to vary both spatially and temporally due to turbulence and granular phenomena such as clustering, bed forms, compaction, grain protrusion, and collective particle motion (Gomez, 1991; Kirchner et al., 1990; Schmeeckle et al., 2001; Strom et al., 2004; Ancey et al., 2008; Zimmermann et al., 2010; Marquis and Roy, 2012; Heyman et al., 2013). This combination of granular and hydrological factors can result in measurements of bed load transport for a single stream that vary by orders of magnitude in time and space (*Recking et al.*, 2012). The understanding of bed load transport in nature is further complicated by intermittency and stochasticity of floods, and with sediment supply that varies in both time and space (e.g., Beven, 2001; Solyom and Tucker, 2004). These complexities at first seem intractable, when presented together with the large degree of temporal and spatial variability in a single catchment or river reach, let alone a landscape or mountain range.

Despite the tumultuous nature of the natural environment, the physical dynamics of rivers and landscapes lend themselves remarkably well to laboratory experiments (*Paola et al.*, 2009). In a sense, heading indoors allows us to better understand what is taking place outdoors. Experiments have provided clarifying demonstrations of how braided rivers interact with vegetation to produce a meandering channel (*Gran and Paola*, 2001; *Tal and Paola*, 2007), or how reducing the supply of coarse sediment to a gravel stream causes it to narrow (*Dietrich et al.*, 1989). Observations from experiments on a one-meter delta have revealed channel dynamics that are directly scalable to natural river systems (*JeroImack and Mohrig*, 2007; *Reitz et al.*, 2010). A recent series of detailed experiments have applied a momentum-conservation framework to predict the dynamics of bed load transport in both laminar and turbulent flows (*Charru et al.*, 2004; *Lajeunesse et al.*, 2010); until this thesis, this framework had not been directly tested in the field. These are but a few examples where laboratory experiments present a clear framework from which to parse the inherent variability in natural systems.

Applying experimental results to the natural landscape is not without challenge. For the case of gravel-bedded-river dynamics, natural systems are driven by highly stochastic forcing, exhibit strong heterogeneity in particle size and shape, and often have dynamics constrained by geological conditions such as bedrock. Connecting results from highly idealized laboratory experiments to the field requires making contact with physical theory, and using this theory to make falsifiable predictions for field systems. At first glance, the limitation of steady flow conditions for bed load transport experiments appears to be a severe oversimplification. Natural floods and sequences of floods are highly transient and intermittent. This may be less of an oversimplification than it seems, however; the seminal work of *Wolman and Miller* (1960) found that, despite catastrophic floods' propensity to rearrange a channel, the majority of rivers are adjusted such that their geometry reflects much smaller floods with recurrence intervals of one to two years. They showed that this more common flood is that whose combination of frequency and magnitude moves the most sediment (*Wolman and Miller*, 1960). Furthermore, it is this flood for which the landscape is adjusted to, rather than the frequent, smallest floods or the infrequent extreme floods (*Wolman and Miller*, 1960). Their work implies that, despite the highly stochastic nature of sediment transport, to understand how a river shapes the landscape one may only need to examine the dynamics of the dominant flood – commonly called the 'bankfull' flood. It has further been demonstrated for gravel-bedded rivers that this flood represents the flow that is at the threshold of motion at the channel banks, and in slight excess of threshold in the channel center (*Parker*, 1978, 1979). Despite large-scale variability in field settings, there is an expectation that gravel rivers exist in a narrowly defined phase-space of applied stress.

It is from these observations that this dissertation appeals to experimental and theoretical evidence, to begin to provide a mechanistic basis for parsing the interaction of climate and mountain landscapes through the entrainment, transport, and deposition of coarse sediment. This dissertation explores several aspects of this interaction: the movement of coarse sediment due to a flood and a series of floods; the long term diffusive behavior of a population of coarse-sediment particles; the implications of threshold channel geometry and the existence of a dominant flood; and the effects of urbanization on channel morphology.

Chapter 2 considers the question of how to quantify variable forcing from floods on bed load transport in a natural river. Considering the displacement of a particle or population of particles under steady flow, we expect that the distance they travel should be proportional to the duration of the flow times the square root of the stress (*Nikora et al.*, 2002; *Lajeunesse et al.*, 2010). In general this expectation holds for a single flood above the threshold of motion, though it is more common to correlate the displacement distance of a particle with the peak or average flood stress (*Church and Hassan*, 1992; *Hodge et al.*, 2011). It is less clear how to quantify the variability of a series of floods, with varying durations and magnitudes. This problem is often circumvented through the application of one-dimensional flow models coupled with sediment transport, however this approach is not easy to generalize or validate. Using dimensional analysis and repeat surveys of tracer particles, Chapter 2 explores the

issue of variable forcing and particle displacement, and derives a novel parameter – the dimensionless impulse – to characterize time-varying floods. We observe the displacement of a plume of tagged cobbles in a flashy tropical river in Northeast Puerto Rico, and compare these measurements to the dimensionless impulse. Results demonstrate that theoretical models for the advection and dispersion of bed load – derived from steady flow assumptions based on laboratory experimental results – quantitatively predict the behavior of bed load transport in a field setting. This framework allows us to infer the distributions of tracer displacement and resting times, and shows that hydrologic forcing may be simplified for long-term studies of sediment transport and landscape evolution.

Even during a flood that exceeds the threshold of motion, individual particles on the river bed spend the majority of their time at rest. When a particle is entrained into bed load, its motion consists of rolling, sliding, and short hops called saltations (Drake et al., 1988). Laboratory experiments show that the number of moving particles scales with the excess dimensionless stress, resulting in two regimes for bed load transport: partial bed load transport where only part of the river bed is mobile at any given time, and continuous bed load transport where the entire bed is mobilized (Wilcock and McArdell, 1993; Haschenburger and Wilcock, 2003). Furthermore, laboratory experiments indicate that, as particles move downstream, the largest particles are preferentially deposited; this eventually leads to the near-universally observed pattern of downstream fining in gravel rivers (*Paola et al.*, 1992b; Paola and Seal, 1995; Fedele and Paola, 2007). Chapter 3 explores the simple question of how a population of bed load particles moves during a single flood, and a sequence of floods of varying durations and magnitudes. This question is explored through the repeated surveying of several populations of tracer particles in the same river as Chapter 2. Again, we find that theories derived from simplistic conditions of laboratory experiments are able to quantitatively predict patterns of transport and sorting in a variable field setting - as long as the hydrograph is properly characterized.

Chapter 4 explores the statistical scaling of floods through the lens of sediment transport.

Climate in rivers is often represented as a distribution of stochastic precipitation events (*Tucker and Bras*, 2000), though this description requires some assumptions as to how hillslopes and gullies alter or pass the signal to the channel (e.g., Tucker, 2004; DiBiase and Whipple, 2011). These assumptions can at times be limiting (e.g., Laque, 2014), and are often circumvented by linking climatic forcing directly to the distribution of floods in the river (Laque et al., 2005). Determining the statistical scaling of floods requires some extrapolation, as recurrence intervals for the largest events are longer than observational records. Observations of self similarity within a flood, and from flood to flood, have prompted the use of power-law distributions (*Turcotte and Greene*, 1993), which seem to reasonably characterize many modern records and are consistent with observations of paleo-flood deposits from the Holocene (*Ely et al.*, 1993; *Knox*, 1993). Landscape evolution models typically parametrize sediment transport rate as a function of river discharge; these models have shown that different statistical descriptions of discharge can result in very different landscapes (Tucker and Hancock, 2010; DiBiase and Whipple, 2011). In particular, researchers have argued – based on the observation of power-law scaling of river discharge – that the importance of extreme events has been overlooked (*Peizhen et al.*, 2001; *Molnar*, 2004). This view suggests that the dynamics associated with the bankfull flood are not the dynamics that are important in shaping the landscape at long timescales (Molnar, 2004). Chapter 4 explores a mechanistic basis for how the statistical distribution of discharge and stress are related to the channel and sediment transport, for a range of gravel-bedded alluvial and bedrock rivers. The analysis in Chapter 4 explores a single catchment with extreme discharge fluctuations in detail, and tests the generality of these observations against a compilation of stream flow records from across the United States. Results show that the threshold of motion acts as a filter on extreme events in precipitation. While discharge may be power-law distributed, the distribution of fluid stress above the threshold of motion is exponential in nearly all streams examined. Stress, not discharge, is the relevant parameter for moving sediment; transport models that parametrize bed load transport as a function of discharge exaggerate the importance of extreme discharge events. The organization of gravel-bedded river channels to near threshold conditions through bed load transport acts to modulate the climate signal; fluid stresses significantly above threshold are exceedingly rare.

The processes which alter a landscape, and thus affect how precipitation is delivered to the river, can take many forms, but perhaps none is more pervasive then the application of impervious surfaces during urbanization (*Chin*, 2006). Compared to timescales of climatic change, urbanization is a geologically rapid process that results in reduced lag times between precipitation and flooding, and increased peak flows (Leopold, 1968, 1991). In general the process of urbanization results in increased sediment delivered to the channel during construction, shifting to a decrease in sediment and larger flood peaks as the landscape is covered in impervious surfaces (Wolman, 1967). These observations are taken from studies in the Eastern United States and the United Kingdom (Chin, 2006). How closely other climatic regions follow this framework is not well established (*Chin*, 2006). Chapter 5 attempts to addresses the question of urbanization's effects on river morphology for tropical regions, by contrasting field surveys of gravel-bedded river channel morphology in a pristine rain forest with channel morphology from high density urban areas. Results show again how the threshold of motion exerts a first-order control on channel form and response to perturbations in coarse-grained channels. Both urbanized and pristine rivers are adjusted to near-threshold conditions; because the tropical environment is naturally strongly flashy – even in the absence of urbanization – we find little influence of impervious cover on channel geometry.

A unifying theme in this thesis is that landscapes are composed of particles, and that the evolution of landscapes ultimately occurs through moving sediment. By focusing on how a flood moves coarse sediment, this thesis attempts to provide a mechanistic basis for the interaction of floods and mountain rivers; connecting the particle scale to the landscape scale. Understanding how the signals of external forcing propagate through rivers may allow us to explore how mountainous landscapes have evolved, and – equally important –

how the mountains might respond to future changes.

CHAPTER 2 : Impulse framework for unsteady flows reveals superdiffusive bed load transport

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Abstract:

Sediment transport is an intrinsically stochastic process, and measurement of bed load in the environment is further complicated by the unsteady nature of river flooding. Here we present a methodology for analyzing sediment tracer data with unsteady forcing. We define a dimensionless impulse by integrating the cumulative excess shear velocity for the duration of measurement, normalized by grain size. We analyze the dispersion of a plume of cobble tracers in a very flashy stream over two years. The mean and variance of transport distance collapse onto well-defined linear and power-law relations, respectively, when plotted against cumulative dimensionless impulse. Data suggest that the asymptotic limit of bed load tracer dispersion is superdiffusive, in line with a broad class of geophysical flows exhibiting strong directional asymmetry (advection), thin-tailed step lengths and heavy-tailed waiting times. The impulse framework justifies the use of quasi-steady flow approximations for long-term river evolution modeling.

2.1. Introduction

Coarse-grained river cobbles spend only a small fraction of their time in motion. Even when fluid stress is above the threshold of motion, a cobble is predominantly at rest under most bed-load transport conditions. Sediment transport at the particle scale can be described as a series of steps and rests, whose respective lengths and durations may be determined by theory or experiment (*Einstein*, 1937). Recent particle tracking experiments along these lines have suggested that bed load particles are separated into mobile and immobile populations, with an exchange rate among them determined by transition probabilities (Ancey et al., 2008). A momentum balance approach at the grain and bulk scales has been utilized to derive relations between fluid shear velocity and particle step length, particle velocity, and number of mobile particles (Charru et al., 2004; Lajeunesse et al., 2010). Video particle tracking in both the field and the laboratory has allowed the determination of tracer diffusion regimes (Nikora et al., 2002) and the physical processes responsible for particle dispersion (Martin et al., 2012). Diffusive regimes are commonly determined through scaling of the diffusion exponent (γ), which relates the variance in particle displacement (σ^2) to time (t) such that $\sigma^2 \sim t^{\gamma}$ (e.g., Metzler and Klafter, 2000). For $\gamma = 1$ diffusion is normal, for all other values of γ diffusion is anomalous where for $\gamma < 1$ processes are sub-diffusive. and for $\gamma > 1$ processes are super diffusive (e.g., Metzler and Klafter, 2000). Nikora et al. (2002) identified three bed-load scaling regimes: the local range of individual ballistic particle trajectories, an intermediate range consisting of many ballistic particle trajectories, and a global range consisting of many intermediate particle trajectories. Martin et al. [2012] showed that dispersion in the local regime is ballistic due to correlated particle motions, and that heavy-tailed particle waiting times caused by burial and scour under low-stage transport could possibly explain anomalous diffusion in the global regime. In the global range both sub and super diffusion have been reported (Nikora et al., 2002; Bradley et al., 2010).

Due to the limited length of laboratory experiments and the demand to understand bed load

in natural systems, particle tracking in the field has become an attractive approach. Sediment tagged with Radio Frequency Identification Passive Integrated Transponders (RFID PIT) allows the tracking of individual particles at long timescales. This tracking method has been used in rivers to determine: particle vertical mixing rates, bed and bed form mobility, virtual velocity, sand and gravel dispersion, and gravel step lengths and rest durations as well as the effects of alluviation and topography on these parameters (e.g., *Habersack*, 2001; Nikora et al., 2002; Ferguson et al., 2002; Haschenburger and Wilcock, 2003; Bradley et al., 2010; Bradley and Tucker, 2012; Hodge et al., 2011). At the flood and multi-flood scale, sediment tracer data represent a cumulative measure of individual particle path lengths (global regime of Nikora et al. (2002)). Given the difficulties of unsteady forcing in a single flood and a series of floods (Figure 2.1(a)), a framework incorporating unsteady flow is essential. Inspired by the impulse framework introduced by *Diplas et al.* (2008) to account for turbulence, we develop a nondimensional impulse to characterize macroscopic variations in fluid stress due to an unsteady hydrograph. We then apply this framework to a new bed load tracer study, and demonstrate that the data collapse onto physically meaningful curves of travel distance and dispersion.

2.2. Impulse Framework

Particles resting on a stream bed require a stress above a threshold value to begin moving (e.g., Buffington and Montgomery, 1997). Laboratory experiments have established that, once a sediment particle is pried free of the bed, it exhibits a velocity and step length linearly proportional to excess shear velocity $(U_* - U_{*c})$, where U_* [m/s] is the shear velocity and U_{*c} is the threshold shear velocity for initiation of motion (Francis, 1973; Fernandez Luque and Van Beek, 1976; Lajeunesse et al., 2010; Martin et al., 2012). The duration of motion for a particle during a flood is unknown; however, it seems clear based on the arguments above that particle displacement should be proportional to the product of (i) the velocity of the particle when in motion and (ii) the duration of flow in excess of the threshold of motion. This can be encoded in a cumulative dimensionless impulse, I_* , that is a kind of transport length:

$$I_* = \int_{t_s}^{t_f} (U_* - U_{*c}) \,\mathrm{d}t / D_{50}, \quad U_* > U_{*c}$$
(2.1)

where t_s and t_f are the starting and finishing times of the hydrograph record considered, and D_{50} is the median grain size of the stream. We consider only flows in excess of U_{*c} in 2.1, which truncates the frequency-magnitude distribution of U_* by excluding all sub-threshold flows (which are irrelevant for transport) present in the hydrograph (Figure 2.1(b)). This removes the effect of flood recurrence intervals which are determined by regional climate and only considers sediment transport as a function of excess momentum imparted on the grains. Our treatment assumes a constant U_{*c} and thus ignores potential variations (temporal and spatial) in the threshold of motion; however, these could be incorporated in the future. For our field study U_{*c} is empirically determined from tracer particles (see 2.3), but a first approximation could be derived from the Shields curve. In the following paragraphs we demonstrate the utility of this dimensionless impulse framework with a field dataset of tracer particles.

2.3. Field Study

The field tracer experiment was performed in the Mameyes River in the Luquillo Critical Zone Observatory in North East Puerto Rico. The Mameyes River is nestled in the heart of the Luquillo Mountains, which have a strong orographic effect resulting in greater than 4000 mm/yr of precipitation. Precipitation occurs as short-duration, high-magnitude events which result in frequent flash flooding (*Schellekens et al.*, 2004). The study reach is a cobble-bedded, 1.2-km stretch of the Mameyes River just downstream of its exit from the mountains; it exhibits nearly uniform width (20 m), minimal meandering, and a slope of $S = 7.8 \times 10^{-3}$. The slope represents a linear regression of the channel longitudinal profile extracted from a lidar DEM (1-m horizontal and vertical resolution). Stage (*h*) was recorded every 5 minutes at the site for 40 days by an In-Situ Level Troll 500, and correlated to measurements from a USGS gage 3.5 km upstream (15 minute resolution) to obtain a stage record for the duration of study.

For the study reach $U_* = (ghS)^{1/2}$ was estimated assuming steady and uniform flow, where g is acceleration due to gravity; Shields stress $\tau_* = U_*^2/(RgD_{50})$, was also estimated for comparison to other studies, where R = 1.65 is the submerged specific gravity of the tracers. We do not attempt to justify the normal flow approximation physically, although channel geometry is remarkably consistent along the study reach; rather, it is a convenient simplification that will be assessed *a posteriori*. We computed the I_* 2.1 for each flood, over the two-year study period (Figure 2.1(a)). We also examined the frequency-magnitude distribution of shear velocity values in excess of critical ($U_* > U_{*c}$), finding an exponential distribution (Figure 2.1(b)); a similar result was obtained when all 20 years of stage data were used. This implies that there is a well-defined average or 'characteristic' shear velocity associated with floods; we return to the implications of this finding below.

RFID PIT tags with unique numbers were installed in 300 cobbles in two separate populations of 150 cobbles placed in the stream, in a 20 by 20 m grid with one meter spacing spanning the channel, in the summers of 2010 and 2011. Tracer particle positions were surveyed two, three, and one time(s) in the summers of 2010, 2011, and 2012 respectively (Figure 2.2(a-c) respectively). Positions were transformed from Cartesian coordinates to a stream-wise normal coordinate system following a methodology similar to Legleiter and Kyriakidis (2007). Total tracer recovery percentages for all six surveys for population one were 62%, 92.5%, 86.6%, 88%, 86.6%, and 93%. Tracer recovery for all three surveys for population two were 100%, 99%, and 94.6%. The low recovery of the initial survey of population one was due to limited sampling time between flooding events. Tracer particle populations were selected from the stream bed to have narrow grain size distributions in order to promote equal mobility (Wiberg and Smith, 1987) among the tracers (Figure 2.2(a inset)). The median grain size values for the stream, and tracer populations one and two, were 12, 12, and 13 cm respectively. The stream D_{50} represents the average of three pebble counts (Wolman, 1954). During transport tracers were fully submerged (average $h/D_{50} =$ 7.0). Tracer particles were located with two wands manufactured by Oregon RFID with maximum detection radii of 50 cm and 20 cm. Survey and detection error were set at 45 cm and 1 m for the small and large wand respectively, which is 1.5 times the calculated combined survey and detection error. RFID tags are detectable at burial depths up to 50 cm and 10-20 cm, depending on tag orientation, for the large and small wands, respectively.

2.4. Results and Analysis

Several individual floods were surveyed in the summers of 2010 and 2011, and used to calculate the distributions of particle travel length and also the fraction of mobile particles, f. For single floods each tracers transport distance (X_i) was normalized by its median diameter (D_i) (Figure 2.3(a)), such that X_i/D_i represents the dimensionless transport distance of an individual tracer. Displacement distances of tracers for each flood are well characterized by an exponential distribution (Figure 2.3(a)). Typical distances of individual tracers were a few meters for a given flood, implying very intermittent (rather than continuous) transport.

We anticipate a linear relation between the peak flood Shields stress and the mobile fraction (i.e., $f \sim \tau_*$) based on momentum balance (*Lajeunesse et al.*, 2010), and determined U_{*c} from the intercept of this fit (Figure 2.3(a inset)). We estimated $\tau_{*c} = 0.023$, or $U_{*c} = 0.22$ m/s, for the tracer D_{50} . Note that f < 1 for all floods, implying that continuous bed load transport did not occur for even the largest events; this is consistent with the inference of intermittent transport from travel distances.

We determined scaling of the mean $(\langle X/D \rangle)$ and variance $(\sigma^2 = \langle (X_i/D_i - \langle X/D \rangle)^2 \rangle)$ of dimensionless tracer transport distances as a function of I_* . Following the reasoning of Section 2.2 we anticipate a linear relationship between I_* and $\langle X/D \rangle$. Here the $\langle \rangle$ symbols represent the ensemble average over all particles. When interpreting fitting exponents of the mean and variance, we make the assumption that tracer displacement has reached the asymptotic limit; i.e., transport is not in a transient regime. Due to the limited number of sampling intervals we utilized all permutations of tracer surveys. Values for $\langle X/D \rangle$ collapse onto a reasonably linear relation when plotted against I_* (Figure 2.3(b)). The fitted equation in (Figure 2.3(b)) represents the ratio of the mean values of I_* and $\langle X/D \rangle$. We also find data collapse of the variance onto a power-law relationship when plotted against I_* (Figure 2.3(c)). There is no systematic trend for $\langle X/D \rangle$ and the variance when plotted against real (clock) time. We note that the values of I_* in Figure 2.3(b) and 2.3(c) are sensitive to the value of U_{*c} , but that the form of the scaling relationships are robust.

The linear form of Figure 2.3(b) implies a constant mean tracer virtual velocity. This suggests that, at long timescales, cumulative impulse is proportional to (and dominated by) the duration of time above threshold, i.e., $I_* \sim t$, where t represents the total time above the threshold of motion. Indeed, we recover nearly identical scaling (as in Figures 2.3(b) and 2.3(c)) when the I_* is replaced by $\langle U_* - U_{*c} \rangle t/D_{50}$, because (see Figure 2.3(c inset)) $I_* \approx \langle U_* - U_{*c} \rangle t/D_{50}$ for our study system, where $\langle U_* - U_{*c} \rangle$ is the average value for the distribution of $U_* > U_{*c}$ (Figure 2.1(b)). The parameter $\langle U_* - U_{*c} \rangle t/D_{50}$ is similar to the dimensionless time used by Nikora et al. (2002) for investigating bed load diffusion scaling. We can now interpret the mean square displacement as representing dispersion of tracers as a function of transport time; the scaling exponent, $\gamma = 1.88$, ($\sigma^2 \sim t^{\gamma}$), indicates super-diffusion.

The tracer waiting time distribution is not directly measurable, however there is an expectation based on experiments by *Martin et al.* (2012) for heavy-tailed waiting times. In the case of symmetric random walks, thin-tailed step lengths and heavy-tailed waiting times produce subdiffusive behavior (e.g., *Weeks et al.*, 1996; *Metzler and Klafter*, 2000). However, in the presence of a strong asymmetry (drift), the same distributions can produce superdiffusive scaling (*Weeks et al.*, 1996). In the case of a river where steps occur in only one direction i.e., downstream long waiting times would appear as Levy flights in the upstream direction when viewed from a Lagrangian reference frame centered on $\langle X/D \rangle$ (*Weeks et al.*, 1996). Indeed, our $\langle X/D \rangle$ measurements support a constant, unidirectional drift (Figure 2.3(b)). Using the analytical framework for asymmetric random walks (*Weeks et al.*, 1996; *Weeks and Swinney*, 1998) and the knowledge that particle step lengths are thin tailed (exponential), we infer that cobbles exhibit a power-law waiting time distribution with an exponent $\nu = 4 - \gamma = 2.12$ for the probability density function. It should be noted that this is the inferred waiting time distribution during flows in above threshold, and that the time below threshold is not considered.

2.5. Discussion and Summary

The novelty and utility of I_* is that it allows the coupling of hydrological and sediment tracer data well beyond the single-flood scale, in a physically-based manner. For channels where the distribution of $U_* > U_{*c}$ is thin tailed, I_* can also be viewed as a dimensionless time following the reasoning of Nikora et al. (2002). An assumption in this study is that U_{*c} is constant, which is not valid in some situations (Kirchner et al., 1990; Charru et al., 2004; Marguis and Roy, 2012). This assumption is made out of necessity as there is currently no feasible manner in which to determine U_{*c} for each flood. However, the collapse of the mean and variance of tracer displacement data suggests that the range of values for U_{*c} cannot be large in this reach during the study period. Our tracer recovery rates (excluding the first flood) were consistently higher than other similar studies (*Ferguson and Hoey*, 2002; Haschenburger, 2011a; Libault et al., 2012). However, there is potential for the scaling exponents γ and ν to be biased by unrecovered tracers, which would likely have traveled the farthest and hence have had significant influence on calculated means and variances. It is possible that tracers were destroyed or buried beyond the detection limit, though the latter is unlikely as similar studies reported fairly shallow burial depths for a large range of conditions (Haschenburger, 2011b; Houbrechts et al., 2012). However, the missing tracers are likely to affect each survey in a similar manner, in that increasing $\langle X/D \rangle$ and σ^2 would simply re-scale the linear and power law relationships for $\langle X/D \rangle$ and σ^2 , respectively. We thus treat our scaling exponents as estimates of their actual values. In the experiments of Martin et al. (2012) the heavy-tailed waiting time distribution was the result of the time it took to scour down to the depth of the buried tracer. If the inferred heavy-tailed distribution of tracer waiting times in this study is accurate, this lends further support to the mobile and immobile partition (Ancey et al., 2008) of sediment tracers, with a residence time in the immobile phase that is controlled by erosion and deposition of the bed (*Martin* et al., 2012).

The similarity of cumulative impulse and $\langle U_* - U_{*c} \rangle t$ implies that unsteadiness of the hydrograph may be further simplified through use of an intermittency factor I = t/T, where T is the total duration of elapsed (clock) time. The time-integrated hydrograph then reduces to a form $I\langle U_* - U_{*c}\rangle T$, which is precisely the treatment used in long-term modeling of river profiles (e.g., Paola et al., 1992a; Parker et al., 1998). For floods exceeding critical shear velocity in our period of study, $\langle U_* \rangle = 0.27$ m/s ($\langle \tau_* \rangle = 0.032$) and the ratio of $\langle \tau_* \rangle / \tau_{*c}$ = 1.39; this is close to the theoretically predicted and measured bankfull-flow values for bed-load dominated alluvial streams at equilibrium, $\langle \tau_* \rangle / \tau_{*c}$ of 1.2 and 1.4, respectively (Parker, 1978; Paola et al., 1992a; Parker et al., 1998). Despite the (perhaps fortuitous) validation of landscape modeling assumptions made by Paola et al. (1992a) this simplification is only valid at timescales sufficiently longer than the recurrence interval of floods and requires that the distribution of $U_* > U_{*c}$ is well behaved (i.e., that it is thin tailed). For the Mameyes, the short recurrence interval of floods exceeding critical (~ 5 days) hastens this convergence. Nonetheless, it is remarkable that the complex hydrograph (Figure 2.1(a)) may be reduced to an average stress with an intermittency factor, and still describe the mean tracer displacement.

Anomalous diffusion of sediment tracers in the global regime of Nikora et al. (2002) has been suggested by modeling (Ganti et al., 2010) and field studies (Bradley et al., 2010; Libault et al., 2012), though usually due to heavy-tailed step lengths. The existence of heavytailed step lengths in natural tracer studies is still a matter of debate (Hassan et al., 2013). In this tracer study we do not observe heavy-tailed step lengths. It is possible that our selection of a narrow tracer grain-size distribution precludes the emergence of heavy-tailed step lengths observed in experiments by Hill et al. (2010). This interpretation is supported by the experimental results of Roseberry et al. (2012), which show thin-tailed step length distributions for nearly unimodal sediment under steady flow. Our inferred heavy-tailed

waiting times emerge only when considering flows above the threshold of motion, and thus are unrelated to flood recurrence (Zhang et al., 2012). Nikora et al. (2002) suggested that heavy-tailed waiting times would lead to subdiffusion in the global regime; however, because sediment tracers undergo asymmetric random walks, the heavy-tailed waiting times produce superdiffusion. Zhang et al. (2012) point out that, at the longest timescales, one should eventually observe normal diffusion because particle waiting times will not be infinitely long; however, in practice the waiting times of field tracers may be sufficiently long that this 'normal' scaling is never observed. At present, we have no basis for estimating the maximum particle waiting time in a river. An additional caveat is that our particles may not have sampled sufficient space and time to reach the asymptotic scaling limit, and so caution should be applied when inferring scaling exponents from these data. If results are taken at face value, however, they suggest that bed load tracers behave similarly to tracers in other geophysical flows (Weeks et al., 1996) and also to charge carriers in amorphous materials (Scher and Montroll, 1975), where heterogeneity leads to long particle trapping times in the presence of strong drift. Furthermore, the dimensionless impulse could act as a catalyst for synthesizing existing bed-load tracer datasets. Should the results of our study be borne out in other rivers, we might be emboldened to extrapolate bed-load dynamics beyond tracer observations using existing hydrologic gage data. If the characteristic flood magnitude and intermittency of a given river could be assumed reasonably constant, one could even estimate bed load travel distances over geologic timescales. This could provide one way to estimate the residence time of pebbles in a river, if their provenance was known.



Figure 2.1: Hydrograph in stage (m) for the Mameyes River for the duration of the field study. The dashed red line represents the empirically determined threshold of motion $U_{*c}=0.22$ m/s (Shields stress of 0.023). The inset is of a single flood with the shaded region representing the flood impulse (see text 2.2). (b) Distribution of shear velocity greater than the critical shear velocity.



Figure 2.2: (a) Location of tracer particle initial placement on lidar DEM of the study area in May of 2010. The inset shows the cumulative grain size distributions for the stream (black line), and tracer population one and two (red and dashed red lines respectively). (b) Location of tracer particles in summer 2011. (c) Location of tracer particles in summer 2012.



Figure 2.3: (a) Dimensionless step length distributions for individual floods normalized by the mean $(\langle X/D \rangle)$ step length for each flood. Dimensionless mean step lengths for each flood are labeled in the legend. The dashed black line represents an exponential distribution. The inset represents the fraction of tracers that moved against the peak Shields stress for each flood. Symbols and colors correspond to the same datasets for both the inset and 2.3(a). (b) Scaling of the mean dimensionless tracer transport distance against dimensionless impulse, the solid black line represents a linear fit through the origin. (c). Scaling of the variance with dimensionless impulse, the solid black line represents the best-fit relationship. The inset shows the equivalence between the dimensionless impulse, and the dimensionless time similar to Nikora et al. (2002).

CHAPTER 3 : Dynamics and mechanics of bed load tracer particles

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Abstract:

Understanding the mechanics of bed load at the flood scale is necessary to link hydrology to landscape evolution. Here we report on observations of the transport of coarse sediment tracer particles in a cobble-bedded alluvial river and a step-pool tributary, at the individual flood and multi-annual timescales. Tracer particle data for each survey are composed of measured displacement lengths for individual particles, and the number of tagged particles mobilized. For single floods we find that: measured tracer particle displacement lengths are exponentially distributed; the number of mobile particles increases linearly with peak flood Shields stress, indicating partial bed load transport for all observed floods; and modal displacement lengths scale linearly with excess shear velocity. These findings provide quantitative field support for a recently proposed modeling framework based on momentum conservation at the grain scale. Tracer displacement shows a weak correlation with particle size at the individual flood scale, however cumulative travel distance begins to show an inverse relation to grain size when measured over many transport events. The observed spatial sorting of tracers approaches that of the river bed, and is consistent with size-selective deposition models and laboratory experiments. Tracer displacement data for the step-pool and alluvial channels collapse onto a single curve – despite more than an order of magnitude difference in channel slope – when variations of critical Shields stress and flow resistance between the two are accounted for. Results show how bed load dynamics may be predicted from a record of river stage, providing a direct link between climate and sediment transport.
3.1. Introduction

Understanding landscape denudation and its relation to climate requires an understanding of how a flood hydrograph drives the sediment mass flux leaving the system through rivers. While suspended sediment represents the largest fraction of mass exiting the landscape (Milliman and Syvitski, 1992; Willenbring et al., 2013), it is coarse bed load transport that sets the limiting rate of landscape incision through its control on bedrock erosion and channel geometry in gravel rivers (Sklar and Dietrich, 2004; Snyder et al., 2003; Parker et al., 2007). Unfortunately bed load transport is notoriously difficult to accurately measure (e.g., Gray et al., 2010) and predict in gravel rivers (Recking et al., 2012). The rate of bed load transport is known to vary both spatially and temporally due to turbulence and granular phenomena such as clustering, bed forms, compaction, grain protrusion, and collective motion (Gomez, 1991; Kirchner et al., 1990; Schmeeckle et al., 2001; Strom et al., 2004; Ancey et al., 2008; Zimmermann et al., 2010; Marquis and Roy, 2012; Heyman et al., 2013), which makes predictions difficult and point measurements highly variable. Bed load is especially difficult to predict near the threshold of motion (*Recking et al.*, 2012), where the dominant transport regime is partial bed load transport, in which only a fraction of the bed is actively mobile at any time during a transporting event (Wilcock and McArdell, 1997). Further confounding this issue is that gravel streams are well known to adjust their geometry to an effective discharge (Wolman and Miller, 1960), which occurs at a flow slightly above the threshold of motion for the median grain size (Parker, 1978; Parker et al., 2007); indicating that partial transport is the dominant transport regime within gravel rivers. The spatially variable and highly intermittent flux during partial transport (Wilcock and McArdell, 1997; Haschenburger and Wilcock, 2003), compounded with the added difficulty of a varying sediment supply, necessitates long term observations to decipher bed load dynamics in the field.

Tracer particles, in particular Passive Integrated Transponder Radio Frequency Identification (PIT RFID) tagged particles, are becoming an attractive low cost and low maintenance method of measuring long term particle dynamics. The application of tracer particles has taken various forms such as: exotic lithologies (*Houbrechts et al.*, 2011), painted bed material (*Wilcock*, 1997a), magnetic (*Hassan et al.*, 1991), radioactive (*Sayre and Hubbell*, 1965; *Bradley et al.*, 2010), and RFID (*Lamarre et al.*, 2005; *Bradley and Tucker*, 2012; *Phillips et al.*, 2013). An advantage of tracer particles is that their installation requires minimal alteration of the stream bed, whereas long term automated bed load measurements often requires the complete reorganization of the channel (*Gray et al.*, 2010). A benefit of RFID equipped tracer particles is that each particle is uniquely identified, which allows its position to be measured at longer time scales with high recovery rates (*Bradley and Tucker*, 2012; *Phillips et al.*, 2013). An advantage of long term observations of tracer particles is that they sample over temporal variations in fluid stress and spatial heterogeneity in the river bed, and thus present an integrated picture of bed load transport dynamics.

In this paper we present the results of a two year deployment of several populations of RFID tracer cobbles within an alluvial river, for flood and yearly timescales. At the individual flood scale we examine the tracer displacement distributions, fraction of tracers mobilized, and mobility of tracers. We show that tracer displacements and the fraction mobile are consistent with results from a recent momentum conservation framework (*Lajeunesse et al.*, 2010). We employ a recently developed dimensionless impulse framework (*Phillips et al.*, 2013) to account for unsteadiness in the hydrograph, which allows us to apply a momentum conservation approach to long-term tracer displacement data. We demonstrate that for flows within the partial transport regime, tracer displacements are close to the limit of one step per flood. Furthermore, we confirm the generality of the long-term tracer displacement results within the main channel with a smaller deployment of tracers in a step-pool tributary. We show that, by accounting for flow resistance and differences in the threshold of motion, displacement dynamics in the step-pool and alluvial channels are similar. Lastly, we analyze and compare the sorting of tracer particles with that of the river to show that the emerging sorting patterns are consistent with a size-selective transport sorting model.

3.2. Theory

In the following sections we present the relevant theoretical background that guides the analysis and interpretations of our tracer particle results. The theoretical background is intended as an introduction to the topics of sediment transport mechanics and dynamics, channel geometry, quantifying hydrologic forcing, and the downstream sorting of sediment by particle size.

3.2.1. Sediment transport at the particle scale

Under a wide range of bed load transport conditions, coarse sediment particles undergo short steps separated by longer periods of rest, which leads to probabilistic descriptions of particle motion (*Einstein*, 1937). A particle step is defined as the distance the particle is transported from entrainment to deposition, and the rest duration is the time between deposition and subsequent entrainment. The motion of particles in bed load transport is comprised of sliding, rolling, or short hops called saltations (Drake et al., 1988), where the travel time is generally much smaller than the rest duration (Lajeunesse et al., 2010; Martin et al., 2012; Furbish et al., 2012a; Roseberry et al., 2012; Furbish et al., 2012b). For near-threshold bed load transport, in which only bed surface particles are mobile, bed load flux may be described as the product of the particle velocity and number density of moving grains (Bridge and Dominic, 1984; Wiberg and Smith, 1989; Parker et al., 2003; Lajeunesse et al., 2010; Furbish et al., 2012a), or similarly the product of the particle entrainment rate and the average particle step length (Einstein, 1950; Wilcock, 1997b; Wong et al., 2007; Ganti et al., 2010; Furbish et al., 2012a). The combination of particle velocity, number of mobile surface particles, depth of the mobile layer, and a threshold stress typically result in a nonlinear relationship between bed load flux and the fluid shear stress (Meyer-Petter and Muller, 1948; Fernandez Luque and Van Beek, 1976; Wong and Parker, 2006; Furbish et al., 2012a). Here it should be noted that the above formulations for bed load particle flux both require averaging the measured quantities over yet undetermined timescales (Ancey, 2010; Furbish et al., 2012a). For steady turbulent flows in the laboratory, the particle velocity and step length have been shown to scale linearly with the excess shear velocity $(U_* - U_{*c})$ (*Fernandez Luque and Van Beek*, 1976; *Lajeunesse et al.*, 2010; *Roseberry et al.*, 2012; *Martin et al.*, 2012), where U_* is the shear velocity (m/s) and U_{*c} (m/s) is the threshold shear velocity for initiation of sediment motion. Specifically, *Lajeunesse et al.* (2010) found that the modal particle step length scales as:

$$X/D = C(U_* - U_{*c})/V_s \tag{3.1}$$

where X is the transport distance (m), D is the particle median axis (m), C=70 is an empirically determined constant, and $V_s = \sqrt{RgD}$ is the settling velocity in the limit of large particle Reynolds numbers, where R is the submerged specific gravity of the particles and g is the acceleration due to gravity (m/s²). The surface density of moving particles was found to increase linearly with the Shields stress (τ_*) (Lajeunesse et al., 2010), where $\tau_* = \tau_b/(\rho_s - \rho)gD$, τ_b is the basal shear stress, ρ_s is the sediment density (2650 kg/m³), and ρ is the fluid density (1000 kg/m³). The dependencies of the step length, particle velocity, and mobile surface density on shear velocity have been recently validated for both unimodal and bimodal grain size distributions under turbulent flow (*Lajeunesse et al.*, 2010; *Houssais* and *Lajeunesse*, 2012). That the recent particle scale framework holds in both laminar (*Charru et al.*, 2004) and turbulent (*Lajeunesse et al.*, 2010) flow encourages us to extend these results to interpret tracer particle data at the field scale.

Treating the particle behavior probabilistically, we focus on the distributions of particle steps and rests. In the laboratory, the distribution of particle step lengths for a given stress and grain size have been observed to follow exponential or gamma-like distributions (*Lajeunesse et al.*, 2010; *Martin et al.*, 2012; *Roseberry et al.*, 2012). However, for mixed grain size distributions, heavy-tailed statistics can emerge due to a summation of exponential step lengths for each size group (*Hill et al.*, 2010). Examining passive tracers in the field introduces an ambiguity; one measures particle displacement – i.e., the distance a particle travels between successive surveys of its position – but this displacement is composed of

an unknown number of steps and rests. Displacement length distributions measured for individual floods, and at longer timescales over many floods, typically follow exponential or gamma-like distributions (*Hassan et al.*, 1991; *Schmidt and Ergenzinger*, 1992; *Habersack*, 2001; *Lamarre and Roy*, 2008; *Bradley and Tucker*, 2012; **?**; *Phillips et al.*, 2013). We propose two simple limits for particle displacement during a flood: (1) the lower limit is that a particle executes a single step, with a characteristic lengthscale predicted by Equation 1; and (2) the upper limit is continuous particle transport, with no rests, for the duration of the flow that exceeds the threshold entrainment stress. We explore tracer displacements within the context of these two limits.

Upon deposition, the rest duration before subsequent entrainment is constrained by two criteria: first the stress must exceed the threshold of motion locally, and second the particle must be exposed to the flow (*Martin et al.*, 2012). The stochastic erosion and deposition of the river bed surface acts to bury and excavate particles, and recent laboratory results suggest that this produces heavy-tailed particle rest durations (*Martin et al.*, 2012; *Martin*, 2013). Although these rest durations cannot be measured from passive tracers in the field, our previous work used the dispersion of the tracer plume to infer similar behavior to laboratory experiments (Phillips et al., 2013). Accordingly, we will not consider the particle rest duration or tracer dispersion in this article.

3.2.2. Dimensionless impulse

Flows in natural coarse-grained rivers are inherently unsteady: from the microscopic scale of variations in turbulence, to macroscopic fluctuations in discharge within a flood, to the rise and fall of the hydrograph throughout a series of floods. At the smallest relevant scales of turbulence, the threshold of motion is determined by the product of shear stress magnitude and duration, the impulse (*Diplas et al.*, 2008). Due in part to the difficulties in measuring tracer particle motion and near-bed stresses during floods, the fluid shear stress is commonly quantified through use of a bulk-flow parameter such as the depth-slope product, $\tau_b = \rho ghS$ (Hassan et al., 1991; Ferguson and Wathen, 1998; Haschenburger and Church, 1998; Church and Hassan, 2002; Lenzi, 2004; Haschenburger, 2011a), where h is the flow depth (m) and S is channel slope. For coarse-grained streams this simplification is perhaps more reasonable, as particle inertial timescales are large and thus coarse particles are insensitive to a range of turbulent stress fluctuations (*Diplas et al.*, 2008; *Celik et al.*, 2010; *Valyrakis et al.*, 2010, 2013). Although some readers may object to the assumption of steady and uniform flow for a flood, the flow may be considered quasi-steady so long as the hydrograph varies slowly compared to the grain inertial timescale (on the order of several seconds), and quasi-uniform so long as water surface slope remains approximately constant. Thus for large ensembles of particles over many floods, the idea of employing a normal flow approximation becomes tenable. Accordingly, *Phillips et al.* (2013) introduced the dimensionless impulse:

$$I_* = \int_{t_s}^{t_f} (U_* - U_{*c}) \,\mathrm{d}t / D_{50}, \quad U_* > U_{*c}$$
(3.2)

to quantify the time-integrated fluid momentum in excess of threshold, assuming normal flow. Here D_{50} represents the median grain size of the tracers, t_s represents the start of a flood, and t_f represents the end of a flood of interest. The integral is only calculated over the record of $U_* > U_{*c}$, as sub-threshold flows do not transport sediment. We found previously that Equation (3.2) reasonably predicts mean particle displacement at annual timescales in the river studied here (*Phillips et al.*, 2013).

3.2.3. Downstream sediment sorting

The spatial pattern of diminishing grain size with distance from the headwaters is near universal among gravel rivers, and results from a combination of size-selective sorting and particle abrasion (e.g., *Paola et al.*, 1992b; *Kodama*, 1994; *Paola and Seal*, 1995; *Ferguson et al.*, 1996; *Gasparini et al.*, 1999, 2004; *Fedele and Paola*, 2007; *Jerolmack and Brzinski III*, 2010). For tracer particles, the relatively short distances and timescales involved preclude abrasion as a mechanism for observed downstream fining (*Ferguson et al.*, 1996). Thus we further explore the mechanisms and implications of size-selective transport as it pertains to

tracer particles and the river bed. We look to laboratory experiments to inform the following analysis, as it is difficult to generalize the rate at which tracer particles sort from previous field studies. Flume experiments with a heterogeneous sediment input show that particles rapidly segregate by size to achieve an equilibrium profile, and that subsequent transport results in an elongation of this sorting profile (Paola et al., 1992b; Paola and Seal, 1995; Seal et al., 1997; Toro-Escobar et al., 2000). The self-similar sorting profile leads to a recasting of the downstream distance into a dimensionless extraction length $X_* = X/L$, composed of the distance downstream X (m) nondimensionlized by the distance L (m) at which 100% of the coarse material in transport has been extracted (deposited). In laboratory experiments and in natural rivers, L is taken as the distance from the input/source to the gravel-sand transition (Paola and Seal, 1995). A value of $X_*=1$ indicates that all gravel particles are deposited upstream of this location. In the absence of a well-defined gravel front the extraction length can be difficult to determine, though Toro-Escobar et al. (2000) suggests that the location where 90 or 95 percent of the source material is extracted by deposition is a suitable proxy for L. In the case of tracer particles, this represents the distance downstream from the source to the location of the 95th percent recovery. Laboratory and modeling results of *Fedele and Paola* (2007) determined that the sorting of a gravel mixture can be described by its variance and mean in the following formulations:

$$\sigma(X_*) = \sigma_o e^{-C_1 X_*} \tag{3.3}$$

$$\bar{D}(X_*) = \bar{D}_o + \sigma_o(C_2/C_1)(e^{-C_1X_*} - 1)$$
(3.4)

where σ_o and \bar{D}_o are the standard deviation and mean of the input material, \bar{D} is the mean grain size, σ is the standard deviation of the grain size distribution, and C_1 and C_2 are constants. For selective deposition in gravel rivers, \bar{D} and σ should decrease exponentially at approximately the same proportion resulting in a constant coefficient of variation (σ/\bar{D}) (*Fedele and Paola*, 2007). These results also suggest that, due to the self-similar sorting profile, separate populations of tracer particles with similar initial grain size distributions (e.g., \overline{D} and σ) should exhibit similar dynamics when properly rescaled by X_* .

3.3. Field site and Methods

Field deployment of tracer particles took place in the Mameyes River basin located in the Luquillo Critical Zone Observatory in North East Puerto Rico. Coarse-grained tracers equipped with PIT RFID tags were deployed in the main stem of the Mameves River, and in a steep tributary called Bisley 3 (Figure 3.1(a)). The Mameyes River sits in the center of the Luquillo Mountains, which possess a strong orographic effect resulting in greater than 4000 mm/yr of precipitation in the headwaters. Precipitation occurs frequently throughout the year in the form of short-duration, high-magnitude events resulting in frequent flash flooding Schellekens et al. (2004). The 1.2 km study reach in the Mameyes begins just downstream of where the river exits the mountains (drainage area of 24.21 km²). Stage (h) was recorded every 5 min at the main channel site for 40 days by an In-Situ Level Troll 500 and measured from surveys of high flow debris following large floods, which were correlated to discharge (Q) measured by a U.S. Geological Survey (USGS) gage 3.5 km upstream (gage #: 50065500, 15 minute resolution) to obtain a reach-average stage record for the duration of the study (Figure 3.2(a)). This section of the Mameyes River (Figure 3.3(a)) exhibits nearly uniform width (20 m), minimal meandering, and has a slope of $S=7.8\times10^{-3}$. The slope represents a linear regression of the channel longitudinal profile (Figure 3.3(b)) extracted from a lidar digital elevation model (DEM) (1 m horizontal and vertical resolution). The smaller headwaters tributary is located in catchment three (drainage area of 0.58 km^2) of the Bisley Experimental Watershed. The Bisley 3 study reach is characterized as a steppool stream with a slope of $S=1.2\times10^{-1}$, width ranging from 2 to 4 m, and boulder steps that range from 0.5 to 2 m in height (Figure 3.3(c)). For this reach (Figure 3.3(c)) the slope was computed as a linear regression of a longitudinal profile surveyed using a laser range finder (Figure 3.3(d)); due to extremely dense forest canopy a longitudinal profile extracted from the lidar DEM does not accurately represent the heterogeneity in channel topography (Figure 3.3(c)). It should be noted that a simple linear regression is unlikely to capture the heterogeneity in transport slopes for this step-pool stream; however we use it to remain consistent with our main channel field site. Stage was recorded every minute at the site for 59 days by an In-Situ Level Troll 500, which was correlated to discharge measured by a U.S. Forest Service (USFS) gage (15 minute resolution) located ~100 m upstream to obtain a reach-average stage record for the duration of the study (Figure 3.2(b)). When calculating the frictional resistance of the stream bed we use the hydraulic radius (h_r) and the D_{84} . We computed h_r as the average hydraulic radius at the threshold of motion from three cross sections surveyed for each site. In order to reduce the variability in the grain size distributions we determined the D_{84} from an exponential fit to pebble count data up and downstream of each site.

Two separate populations of 150 tracers were installed in the summers of 2010 and 2011 in the Mameyes. A smaller population of 51 tracers were installed in the Bisley 3 stream in the summer of 2010. All three tracer particle populations were selected from the stream bed to have narrow grain size distributions centered on the stream D_{50} (Figure 3.1(C-D)). Narrow grain size distributions were selected in order to promote equal mobility (Wiberg and Smith, 1987) within the tracer populations. The median grain size values for both tracer populations and the river bed at the Mameyes site were 12, 13, and 11 cm, respectively. Median particle diameters for the Bisley 3 stream and tracer particles were 12 and 13.5 cm, respectively. The median grain size for the main channel reach represents the average of three pebble counts (Wolman, 1954). In the Mameyes and Bisley 3 field sites the tracers are fully submerged (average at both sites $h/D_{50}=7$) during transport for flows above the threshold of motion. Both populations of tracers in the Mameyes River were deployed in the same reach (Figure 3.1(e)) in a 2020 m grid with 1 meter spacing, spanning the width of the channel. They were surveyed two, three, and one time(s) during the summers of 2010, 2011, and 2012, respectively. Tracer recovery percentages for the six field surveys for the first population were 62%, 92.5%, 86.6%, 88%, 86.6%, and 93%. Recovery percentages for the second tracer population for field surveys in 2011 (2) and 2012 (1) were 100%, 99%, and 94.6%. The low initial recovery rate for population one resulted from an incomplete survey due to a following flood. The second population of tracers was placed in the river as two installments of 80 and 70 tracers on two consecutive days due to a small flood, which resulted in a minor amount of burial from fine sediment for the initial 80 tracers installed. Impacts of the initial increased embeddedness are not observable at the multi-flood scale. Unrecovered tracers have the potential to bias the mean value, as it is possible that tracers could be buried beyond the detection limit, destroyed, missed, or are further downstream. However, it was common to find previously missing tracers on subsequent surveys suggesting that the unrecovered tracers were buried or missed. For all surveys of the first population we cannot account for 7% of the installed tracers. Tracers were surveyed in the Bisley 3 stream two, four, and one time(s) during the summers of 2010, 2011, and 2012, respectively. Tracer recovery percentages for the seven surveys were 91%, 91%, 93%, 98%, 98%, 100%, and 93%. Final surveyed positions of the tracers can be seen in Figures 3.1(b) and 3.1(f). Surveyed positions for both field sites were transformed from Cartesian coordinates to a stream-wise normal coordinate system following a methodology similar to that of *Legleiter* and Kyriakidis (2007). Tracer particles were located using two wands manufactured by Oregon RFID with empirically determined maximum horizontal detection radii of 50 cm and 20 cm respectively. RFID tags were detectable at burial depths up to 10-20 cm below the river bed for the small wand (depending on tag orientation), and 50 cm for the large wand. The maximum combined survey and detection error are estimated to be 45 cm and 1 m for the small and large wands, respectively. All measured tracer motion recorded below the detection threshold was considered to be error and set to zero.

To determine the spatial sorting of the stream bed we performed pebble counts spaced every 200 m downstream of the location of the Mameyes RFID tracers. For the stream sorting we restrict the analysis to the depositional part of the river, and thus we only use measurements downstream of the start of the alluvial plain. Statistical values for the tracers represent the spatial average at the center of eight linearly spaced bins, where the number of bins was determined to ensure enough tracers within each bin for accurate statistics. To determine the dimensionless distance X_* for the stream we set the basin length (L) as the distance from the start of the alluvial plain to the perceived gravel sand transition. The exact location of the gravel sand transition in the Mameyes River is obscured in the field due to substantial anthropogenic modification of the river. However, with the use of airborne lidar data we have identified a substantial break in channel slope within the confines of the golf course approximately 3.5 km downstream of the start of the alluvial plain, which we take as the gravel sand transition. As there is no well defined front for both populations of tracer particles we set L as the distance at which 95% of the tracers recovered remain upstream for each population of tracers. Ideally L is the distance at which 100% of the total (recovered and unrecoverable) tracers remain upstream, however we cannot accurately determine this point.

For the study reaches $U^* = \sqrt{ghS}$ was estimated assuming steady and uniform flow; Shields stress was also estimated for comparison to other studies. Long-term flow records for the Bisley 3 stream are measured near a series of large culverts that artificially truncate the largest floods; this does not effect the calculations of I_* greatly, but does add ambiguity to the distributions examined in following sections. Therefore, we limit our analysis to the hydrograph on the main channel of the Mameyes River, as it represents the highest quality data.

The value of I_* (3.2) was computed for each flood above the threshold of motion (Figure 3.2), and also for the cumulative time periods between successive tracer surveys. The calculation of I_* is particularly sensitive to the value of U_{*c} , a parameter that is known to vary both temporally and spatially (*Kirchner et al.*, 1990; *Charru et al.*, 2004; *Turowski et al.*, 2011; *Marquis and Roy*, 2012). We determined the value of U_{*c} in two manners: one, U_{*c1} was calculated from the fraction of mobile tracers for individual floods, and two, U_{*c2} was determined as the value that provided the best collapse of the mean tracer displacement data (both methods are used in Sections 3.4.2 and 3.4.2). The sensitivity of Equation (3.2) to the determination of U_{*c} is a shortcoming of the dimensionless impulse framework when applied to near-threshold gravel channels; however, the determination of a critical stress for incipient motion in coarse gravel streams remains an outstanding problem in fluvial geomorphology. Using two definitions of critical, the resulting values of I_* for a representative flood are 3815 for U_{*c1} and 806 for U_{*c2} (Figure 3.4). The discrepancy between the values of I_* is due to the broadening asymmetric shape of the hydrograph, where a small change in U_{*c} can result in an order of magnitude increase in I_* . It should be noted that both approaches produce the same scaling relationships, but differ in the magnitude of the coefficients. A potential drawback of calculating U_* and U_{*c} from a reach-averaged depth-slope product is that large instantaneous values of U_* due to turbulent fluctuations cannot be accounted for. This simplification could result in tracer movement from turbulent fluctuations being attributed to a reach-averaged stress, and hence a biased estimate for the actual threshold value. The upshot is that estimates for I_* are least accurate for low magnitude, short-duration floods. Data points likely to be affected by this consequence of determining U_{*c} are included for completeness and indicated in the following figures where appropriate. Due to this drawback we do not recommend using Equation (3.2) for individual floods without an independent measure of U_{*c} . For the remainder of the manuscript, except where noted, we use the value of U_{*c} as determined by the least-squares regression of the fraction of mobile tracers (U_{*c1}) .

3.4. Results

3.4.1. Hydrology filtered through sediment mechanics and dimensionless impulse

River discharge is the most commonly reported variable in sediment transport studies that relate long-term sediment dynamics to hydrologic forcing. For considerations of bed load transport, however, the threshold of motion applies a filter to these data; only flows that exceed the critical stress for entrainment are relevant for assessing particle transport. Moreover, the momentum framework presented above reminds us that fluid stress – rather than water discharge – is the relevant parameter to consider for driving sediment motion. Accordingly, we empirically estimate the threshold of motion by determining the intercept in a plot of the fraction of mobile tracers f against peak flood stress for several individual floods, where tracers' positions were surveyed before and after the event. Based on a momentum balance approach (Lajeunesse et al., 2010) we anticipate that f should scale linearly with Shields stress (Figure 3.5). A linear relation provides a reasonable fit to the data when plotted against the peak Shields stress for the D_{50} of the tracer population. We treat the intercept as the threshold of motion, determining that $\tau_{*c}=0.023$ ($U_{*c}=0.22$ m/s) (Figure 3.5). We use the Shields stress at the flood peak – rather than average Shields stress – as it does not require additional information; computing average stress associated with a flood requires choosing a threshold value for fluid stress. Error bars for f were calculated based on the number of missing tracers, where the upper and lower lines represent the absolute maximum and minimum values for the fraction mobile by assuming that all tracers not recovered moved or didn't move, respectively. For all flood events monitored, the fraction mobile remained well below one (f < 1). With knowledge of the threshold of motion we can readily discern that there were numerous sediment transporting events within the hydrograph timeseries (Figure 3.2(a)).

The frequency-magnitude distribution for U_* for the entire period of record (Figure 3.6(a)) presents a fuller picture of the statistical scaling of flow within the hydrograph. For small to intermediate values of U_* the curve appears to be a straight line on a log-log plot (Figure 3.6(a)), indicating potential power-law scaling for this region, a common feature of flood hydrology (*Turcotte*, 1994; *Lague et al.*, 2005; *Molnar et al.*, 2006). The power-law scaling is even more evident for discharge (Figure 3.6(a. inset)). The frequency-magnitude distribution for both U_* and Q exhibits a truncation to the power-law scaling, that occurs at approximately the threshold of motion (Figure 3.6(a)). The upper truncation of the distribution is well fit by an exponential function, indicating that the shear velocity for flows exceeding threshold is well-described by a single average value, $\langle U_* \rangle = 0.27 \text{ m/s} (\langle \tau_* \rangle = 0.033)$. The exponential decay in probability for shear velocities above critical indicates that most floods are only slightly above the threshold of motion. In terms of shear velocity, the peak value associated with an observed bed load transport event was $U_*=0.46 \text{ m/s}$ ($\tau_*=0.1$). Although the highest discharge values in the 20-yr instantaneous record exceed our peak observed flood by a factor of 3.1 ($Q=444.57 \text{ m}^3$ /s), the highest shear velocity values associ-

ated with these extreme events are only 1.28 ($U_*=0.59 \text{ m/s}$) as high as the peak observed value in this study. This analysis indicates that most bed load transport in the Mameyes occurs close to the threshold of motion.

We analyze the frequency-magnitude distribution of I_* (Figure 3.6(b)) and find that the distribution of I_* is composed of several scaling regions, though it does not appear to be heavy-tailed. The smallest values of I_* are artificially truncated by the resolution of the river stage measurements (15 min). The probability density function (PDF) of I_* (Figure 3.6(b. inset)) has a pronounced peak that coincides with the product of the average excess shear velocity and average flood duration, $(\langle U_* \rangle - U_{*c}) \langle t \rangle / D_{50}$. Here $\langle U_* \rangle$ is the average value of the frequency magnitude distribution of $U_* > U_{*c}$ (Figure 3.6(a)), and $\langle t \rangle$ is the average duration of a flood above the threshold of motion for the period of record. The distributions of U_* and I_* calculated over the relatively short study duration (2 years) are the same as those calculated using a longer flow record (20 years available from the USGS) for the Mameyes River. Thus, data indicate that there is a well-defined 'characteristic flood' for the Mameyes.

3.4.2. Mechanics of sediment tracer particles

Individual flood scale

The distribution of particle displacements resulting from several floods was determined from surveys of both populations of tracers at the Mameyes site, in the summers of 2010 and 2011. Each tracer's transport distance (X_i) was normalized by its median grain diameter (D_i) . For individual flooding events above the threshold of motion, the majority of the cumulative distribution functions (CDF) of tracer particle displacement are well-described by exponential functions (Figure 3.7(a)), except for two tracer displacement CDFs which decay faster than exponential functions (green circles and cyan squares in Figure 3.7(a)). Each tracer displacement CDF is normalized by its mean displacement ($\langle X/D \rangle$) to facilitate plotting all single events in one graph (Figure 3.7(a)). Typical travel distances for individual tracers were on the order of a few meters for each flood. We plot the dimensionless tracer distance for all single events against the peak shear velocity for that event, normalized by V_s , and find that the modal tracer displacement is well described by a linear relation, as anticipated by the momentum framework presented earlier. A fit to the modal displacement distances provides another estimate of the threshold stress, $U_{*c}=0.13$ ($\tau_{*c}=0.016$), from the intercept. This is likely a lower estimate of U_{*c} , and is not the same value as determined previously for the fraction mobile data. Finally, we plot Equation (3.1) over a contour density map of our field data (Figure 3.7(b)), using k=70 from Lajeunesse et al. (2010). Remarkably, the modal step lengths predicted from the laboratory-derived relation of Lajeunesse et al. (Equation 3.1) run through the modal displacement distances measured from our field data in the Mameyes.

Multi-flood scale

At the multi-flood scale we analyze the long-term behavior of the tracer particles' displacement. We normalize the CDFs of cumulative travel distance by each survey's mean value, which results in a collapse of the data. We note here that the CDFs in Figure 8 are truncated at the lower end due to measurement accuracy, and at the upper end of the distribution due to unrecovered tracers (see ? for a discussion of the effects of unrecovered tracers on the scaling of the CDF, while for a discussion of the functional distribution for the CDF see *Bradley and Tucker* (2012)).

At the multi-flood scale we analyze the mean cumulative tracer travel distances $(\langle X/D \rangle)$ using the dimensionless impulse (I_*) (3.2). We find that the $\langle X/D \rangle$ scales linearly with I_* for both populations of tracer particles (Figure 3.9). Despite the second population of tracer particles being less embedded in the stream bed, its $\langle X/D \rangle$ follows the same trend as the first population when plotted against I_* . Due to a limited number of repeat surveys we utilize all permutations of tracer surveys; i.e. we determine the $\langle X/D \rangle$ and I_* for all possible sampling intervals. Using all permutations of tracer surveys does require the assumption that the sequence of floods does not exert substantial control on the mechanics of particle displacement. However, flood sequence and particle embeddedness may explain some of the scatter in the $\langle X/D \rangle$ displacement data (Figure 3.9). Due to the close agreement between our field data and laboratory results (Figure 3.7(b)) we use equation (3.1) to calculate the two limits of particle transport discussed in section 3.2.1. Limit 1, in which entrained particles execute one step per flood, was calculated using equation (3.1) with the peak U_* for each flood for the study duration above U_{*c1} . Limit 2, continuous motion, was calculated as the product of equation (3.1) and t/T_s where t is the duration of the flood above the threshold of motion, and $T_s = 10.6\sqrt{D_{50}/Rg}$ is the expected particle step duration from $Lajeunesse \ et \ al.$ (2010). For limit 2 we use the average shear velocity over the duration of a flood in equation (3.1). When these limits are compared with the tracer particle data, we find that the tracer particles' mean displacement is significantly closer to limit 1 than limit 2 (Figure 3.9), consistent with intermittent bed load transport.

Due to time limitations in the field we were unable to survey more than three individual floods for the Bisley 3 stream, and thus are unable to determine the threshold of motion from the fraction of mobile tracers. In order to compare the Mameyes data with the Bisley 3 tracer data, values for U_{*c} from both sites must be determined using the same methodology (*Wilcock*, 1988). Therefore we utilize method two, described in section 3.3, to determine the values of U_{*c2} for this comparison. The values of U_{*c2} that provide the best collapse of the tracer data are 0.28 m/s ($\tau_{*c2}=0.038$) and 0.81 m/s ($\tau_{*c2}=0.303$) for the Mameyes and Bisley 3 sites, respectively. We find that the Bisley 3 data are also well characterized by a linear relation, with a slope that is one order of magnitude lower than that of the Mameyes site (Figure 3.10). It is intriguing that the Bisley 3 displacement data form a well defined linear trend, considering the limited number of tracers and the rough channel geometry (Figure 3.3(c-d)). We note here that the slope of the linear relation computed with U_{*c2} for the Mameyes site is an order of magnitude larger than that determined using U_{*c1} ; however, the linear form of the relationship is robust for a wide range of threshold values.

3.4.3. Tracer particle sorting in the Mameyes

At longer timescales, tracer particle sorting by size is readily apparent in the Mameyes, as observed in other studies (Hassan et al., 1991). Sorting may be the result of: (1) an inverse relation between particle displacement and grain size; and/or (2) differences in entrainment frequency as a function of grain size. To examine differences in entrainment, we separate the tracer particles into two populations: particles that moved at least once for all single floods, and particles that remained immobile. All distributions are well fit by log-normal functions. We aggregate all of the mobile and immobile particles into two combined distributions for statistical purposes (Figure 3.11(a)). The size distribution of mobile particles is finer at the coarse end of the distribution compared to immobile particles (Figure 3.11(a)), though there is a fair amount of overlap between the CDFs. We used a two sample t-test (equal variance, unequal sample size) on the natural-logtransformed distributions to determine that the difference in the mean values $(\langle mobile \rangle = 119)$ mm, $\langle immobile \rangle = 137$ mm) between the mobile and immobile populations is statistically different (t - statistic = 4.02, degrees freedom = 345, p - value < 0.001). At the singleflood scale there does not appear to be a significant dependence of displacement length on particle size. When we compare the displacements during a flood to the expected step length calculated from equation (3.1); it is clear that large particles have displacements that are close to the expected step length, while a significant number of small particles have much longer displacements (Figure 3.11(b)). This observation, combined with grain-sizedependent entrainment, suggests that the relatively large displacements for small particles result from a greater frequency of entrainment events during a flood. In other words, the largest particles appear to take one step during a flood due to a single entrainment event, while small particles may take multiple steps.

The cumulative effects of minor grain size sorting at the flood scale (Figure 3.11) result in the rapid development of tracer sorting at the annual scale for the Mameyes River. In order to connect the dynamics of tracers to the downstream fining pattern of river bed, we first analyze the downstream grain-size trend of the bed surface of the Mameyes River as determined from pebble counts. The starting location of the tracers coincides with where the Mameyes River exits the mountains (approximately at 0 km downstream in Figure 3.12(a)). The start of the alluvial plain begins at approximately 0.5-0.75 km downstream in Figure 3.12(a). From this point downstream the bed D_{84} fines by roughly a factor of two, while there is only minimal decrease in the bed D_{50} and very slight change in the bed D_{16} (Figure 3.12(a)). Thus, it appears that downstream fining of the river is accomplished by deposition of the coarsest particles. Starting at 300 m downstream the standard deviation and mean decrease downstream exponentially with initial statistics of $\sigma_o=185$ mm and $\bar{D}=185$ mm (Figure 3.12(b & e)). The exponent of the fitted exponential function (C_1 in Equation 3.3) is 1.2 for the stream. This results in an approximately constant coefficient of variation $\sigma/\bar{D}=0.93$ (Figure 3.12(c)). Using the parameters determined in Figures 3.12(b-c) we apply the full model of *Fedele and Paola* (2007) (Equation 3.4) to the spatial decrease in \bar{D} and find that the model seems to under predict the mean value (Figure 3.12(e)). Potential reasons for the underfit are given in section 3.5.2.

Turning to the tracers emplaced in the Mameyes, sorting of both populations by size is readily apparent (Figure 3.12(a)). The spatial decreases σ for both tracer populations (σ_{P1} and σ_{P2}) are well described by exponential functions (Figure 3.12(b)). The exponents of the fitted exponential functions (C_1 in Equation 3.3) for population one, and population two are 0.83 and 0.73, respectively. The initial standard deviation of tracer populations one and two at placement are 48 and 41, respectively. These values are very close to the coefficients for the fitted exponential in Figure 3.12(a). The initial mean values for the tracer populations one and two are 129 and 140, respectively. In accordance with the model of *Fedele and Paola* (2007) we find that both populations of tracers have nearly constant coefficients of variation (σ/\bar{D}) of 0.25 and 0.26, respectively (Figure 3.12(c)). The coefficient of variation for the tracers is expected to be low due to the narrow grain size distribution. Furthermore, we find that the σ data can be reasonably collapsed by normalizing by σ_o from the upstream end of the depositional system (Figure 3.12(d)). Here we apply equation (3.4) using the parameters determined in Figures 3.12(b-c) and find that it under predicts the rate at which \bar{D} decays. In applying equation (3.4) for the tracers we set $\bar{D}_o=173$ to the value for the stream at the location where they were placed. This is because the model is for a bed with a continuous source where the mean at $X_*=0$ does not change, while for a finite population of tracers the coarser particles are deposited and thus over time \bar{D} at $X_*=0$ will coarsen. Potential reasons for the underfit are given in section 3.5.2.

3.5. Discussion

3.5.1. Sediment mechanics

At the single-flood scale, tracer particle displacements have been shown to be well described by exponential, gamma, and power law distributions (*Phillips et al.*, 2013; *Habersack*, 2001; ?). Here we find that the majority of observations are well described by exponential distributions, with two exceptions that decay faster than exponential (Figure 3.7(a)). The distributions that decay faster than exponential are likely due to under sampling as a result of the small number of mobile tracer particles in these floods. The particle velocity distribution has been shown in laboratory experiments and theoretically to scale exponentially, except for small sample sizes for which the distribution decays faster than exponential (Furbish and Schmeeckle, 2013); this pattern is also likely true for the distribution of particle displacements. The remarkable agreement between the modal tracer displacement data at the single-flood scale for the Mameyes, and the laboratory results of Lajeunesse et al. (2010)(Figure 3.7(b)), strongly suggests that the most likely tracer displacement during a flood is a single step length. This holds for all observed individual floods, despite a three-fold increase in Shields stress ($\tau_*/\tau_{*c1} = 3.04$), demonstrating that partial and intermittent bed load transport occurred under all observed conditions. However, tracer displacement-length distributions become increasingly skewed toward larger values with increasing flood strength (Figure 3.7(a & b)), indicating that increasing numbers of particles experience more frequent entrainment as shear stress increases. Nonetheless, the fraction of mobile tracers remained significantly below one for all observed floods (Figure 3.5), reinforcing the inference that

partial bed load transport is the dominant mode of transport here. Finally, plots of mean tracer displacement length against cumulative impulse show that tracer motion is far from the continuous limit, and much closer to the lower limit of one step per flood (Figure 3.9).

The agreement of observed tracer displacement distributions from the Mameves with models and laboratory data (Lajeunesse et al., 2010) gives us hope that the results here are general. Our earlier observations of tracer dispersion provided an indirect method for inferring particle rest times (*Phillips et al.*, 2013): thus, results from this tracer study may provide the basis for future probabilistic modeling of long-term bed load transport, for which the distributions of particle steps and rests are required input parameters (Zhang et al., 2012). As pointed out above, although larger floods have occurred in the historical record of instantaneous discharge, the associated stresses were no more than 1.28 times the largest observed flood. From the USGS records of annual maximum flood peaks (1967-2012, discontinuous) the largest flood is only 1.35 times our observed largest value of U_* . The rapid decay of frequency of occurrence for flows above threshold (Figure 3.6(a)) indicates that transport conditions observed during our study are representative of the river system. Partial bed load transport during near-threshold conditions is also consistent with expectations from equilibrium channel theory for gravel rivers (*Parker*, 1978; *Parker et al.*, 2007). Indeed, the channel depth inferred from the hydrograph for the 'characteristic flood' on the Mameyes - i.e., the peak value in the impulse distribution - agrees with an independent estimate of bankfull flow deduced from vegetation markers, channel morphology, and flow frequency analysis (*Pike*, 2008; *Pike and Scatena*, 2010). More broadly, the tracer dynamics observed in the Mameyes are likely similar in many gravel rivers, because most gravel rivers exhibit near-threshold transport associated with their respective characteristic floods.

We now turn to data from the Bisley 3 tracer deployment, which can serve as a critical test of the generality of the impulse framework and tracer displacement results. Given the small number of tracers deployed and the particularly variable stream profile (Figure 3.3(d)), it is surprising that the Bisley 3 data fall on a well-defined linear relation when plotted against I_* (Figure 3.10). The offset between the two field sites indicates that, for equivalent values of I_* , tracer particles at the Mameyes field site have traveled farther than those in the Bisley 3 stream. This could result from enhanced particle trapping and hiding effects, or greater flow resistance due to the rougher bed. We attempt to collapse the Mameyes and Bisley 3 data onto a single curve by accounting for each of these two effects separately. We assess particle hiding effects using a simple hiding function (*Einstein*, 1950; *Wilcock and Crowe*, 2003), which does not produce a collapse of the data. To test the effect of flow resistance, we calculate the dimensionless friction factor K_f using a modified Keulegan equation that was found to provide the a reasonable fit to a large compilation of field data (*Ferguson*, 2007):

$$U/U_* = \sqrt{8/k_f} = (1/K)ln(11h_r/4D_{84})$$
(3.5)

where U is the flow velocity (m/s), and K=0.41 the Von Karman constant. The h_r for the Mameyes main channel and Bisley 3 sites are 0.9 m and 0.29 m, and the D_{84} is 0.31 m and 0.55 m, respectively. For the Mameyes and Bisley 3 reaches k_f is 0.30 and 9.27, respectively. When values for I_* are normalized by $\sqrt{k_f}$ computed using equation (3.5), the mean tracer displacement data for the Mameyes and Bisley 3 streams collapse onto a single curve (Figure 3.13). This collapse indicates that, when one accounts for the difference in relative submergence and its effect on flow resistance, that the resulting particle transport is the same. The morphologies of these two streams represent end members for the Mameyes watershed, indicating that the linear function $\langle X/D \rangle = 0.025I_*/\sqrt{k_f}$ may be a general relationship. This conjecture would be well supported should this relationship be found to hold for tracer studies in other regions. We note here that one can achieve a similar collapse of the data using other recently proposed flow resistance equations as well (see *Ferguson*, 2007; *Rickenmann and Recking*, 2011; *Ferguson*, 2012).

A pitfall of the dimensionless impulse is its sensitivity to the determination of U_{*c} . As seen in this manuscript, and in general (*Wilcock*, 1988), the value of the threshold stress is dependent on the method used to determine it. In this manuscript we have determined two separate values for U_{*c} by using the intercept of the fraction mobile data ($U_{*c1}=0.22$, $\tau_{*c1}=0.023$, Figure 3.5), and the value that best collapses the long term mean displacement data ($U_{*c2}=0.28$, $\tau_{*c2}=0.038$, Figure 3.10). Both of the determined values are within the range reported from field and laboratory data (*Buffington and Montgomery*, 1997; *Mueller et al.*, 2005; *Lamb et al.*, 2008), however the range of values we've recorded underscores the need for an independent empirical measure of the threshold of motion.

3.5.2. Sediment sorting

Results indicate that the modal displacement length for particles of all sizes is approximately one step, and that the dimensionless particle step length depends only weakly on particle size (Figure 3.11(b)). The latter result appears to support laboratory experiments that show that the largest particles travel the farthest for rivers with steep slopes (Solari and Parker, 2000; Hill et al., 2010). This effect has been attributed to particle inertia for narrow unimodal sediment size distributions (Solari and Parker, 2000), and to bed roughness effects for wider grain-size distributions (Hill et al., 2010). Our field results suggest that sorting happens through smaller particles possessing a higher probability of reentrainment, rather than possessing longer step lengths as compared to larger particles. This may be a consequence of near-threshold transport conditions in the Mameyes, but more work is needed. The pattern of smaller particles having larger displacements (Figure 3.11(b)) is near universally observed in tracer studies (Church and Hassan, 1992; Ferguson and Wathen, 1998; Hodge et al., 2011; Scheingross et al., 2013), though the normalization is often done with the mean displacement length and not an expected step length. For steep streams the particle size dependence on transport distance are similar to what we have observed in the Mameyes (Lenzi, 2004; Scheingross et al., 2013).

The cumulative sorting results over annual timescales appear to substantiate aspects of the self-similar sorting theory of *Fedele and Paola* (2007), and are in general agreement with earlier laboratory experiments (*Paola et al.*, 1992b; *Paola and Seal*, 1995; *Seal et al.*, 1997; *Toro-Escobar et al.*, 2000). The two tracer plumes in the Mameyes had significantly different front positions (Figure 3.12(a)) at the end of the study, as they were emplaced in different years; the first population terminated at X = 1 km while the second ended at X = 1.2 km. However, the two populations behave dynamically similar when their travel distances are rescaled by their extraction lengths (Figure 3.12(b-e)). The curve describing the downstream decay of the standard deviation for tracers is offset from that of the river bed, and this is due to the fact that tagged tracers exhibit a much narrower grain-size distribution than the substrate of the river (Figure 3.1(c)). The scaling exponents (C_1 in Equation 3.3) for the tracers fall well within the predicted range of 0.5-0.9 (Fedele and *Paola*, 2007), however the stream's exponent is larger. Accordingly, we normalize the tracer and river-bed data by their initial standard deviation at the start of the depositional section of the river (X = 0 km); the result is that tracer sorting appears to track that of the river bed when the initial particle size population and distance traveled are taken into account (Figure 3.12(b-e)). Data show that the initial establishment of the sorting profile can be quite rapid, in that the tracer particles behave dynamically similar to the stream despite different residence times within the river. To our knowledge, this tracer study represents the first active field confirmation of the selective deposition theory (*Paola et al.*, 1992b; Paola and Seal, 1995; Seal et al., 1997; Toro-Escobar et al., 2000; Fedele and Paola, 2007)).

The full model predicting the mean concentration of the sediment plume downstream consistently under predicts the data further downstream (Figure 3.12(e)). There are several reasons to expect that a finite population of sediment tracers will not follow the model in equation (3.4). The low values of σ at the leading edge of the tracer plume may have reached a limit where the size differences between tracers is negligible and sorting ceases for this mixture. Where as for the stream the under prediction is potentially from under sampling at the downstream end due to an artificial truncation caused by anthropogenic modification of the stream, which results in a higher value of C_1 in equation (3.3). As the scaling exponent is also fairly sensitive to the determination of L in calculating X_* in this short system. In the case of the stream bed a small shift in L can steepen or elongate the sorting profile resulting in larger or smaller scaling exponents, respectively. Another factor complicating the scaling of the sorting in the Mameyes River is that the distance from the mountains to the ocean is particularly short (5.95 km), resulting in an abrupt truncation of the sorting profile somewhere within the final kilometer of the river. Given the short time that it took for the tracer particles to sort, and the minimal decline in D_{50} downstream one might expect that there should be a rapidly prograding gravel front (*Parker and Cui*, 1998), however this front may be arrested due to Holocene sea level rise *Toscano and Macintyre* (2003).

3.6. Conclusions

In this paper we have presented field results on bed-load tracer displacement data at the event to annual timescales, and used simple theory to: rationalize the displacement scaling of tracer particles; show how a heterogeneous population of particles sorts downstream; and explore the implications of these findings on the statistical scaling of the hydrograph. At the scale of single floods, the distribution of particle displacements is well described by an exponential distribution. Close agreement with laboratory data and theory (Lajeunesse et al., 2010) suggests that these displacements represent the scaling of the fundamental particle step length. We infer that, for near-threshold floods, the most probable transport distance is one step length. Cumulative displacement over many floods reinforces this finding, with data showing that tracers remains in the partial transport regime for a range of flow conditions. We test the applicability of the impulse framework using data from two streams of very different morphologies, and find that tracer displacement data collapse onto a single linear relationship, when differences in critical Shields stress and flow resistance are accounted for. For particle sorting, we find that downstream fining emerges after a series of floods. Sorting seems to results from a slight difference in size-dependent particle entrainment at the flood scale. We find that tracers sort to the limit of sorting present in the stream bed. Both the tracers and the stream bed have the same scaling when accounting for the distance each has traveled, and the initial statistics of the tracer and stream grain size distributions, respectively. These observations serve as an active field

validation of the selective deposition sorting model (*Paola et al.*, 1992b; *Fedele and Paola*, 2007). Finally we show that the frequency-magnitude distribution of flood stress in the Mameyes is exponential for flows exceeding the threshold of motion. The average stress for flows exceeding critical is approximately 1.4 times the critical Shields stress, and represents the stress of maximum geomorphic work. In addition, the distribution of dimensionless impulse has a well defined peak coincident with the flood of maximum geomorphic work, indicating that the channel is adjusted to a characteristic flood impulse. We believe that tracer dynamics observed in the Mameyes River are characteristic of many gravel rivers, because most gravel streams are adjusted such that bankfull floods exert a stress that is only slightly in excess of the threshold for entrainment (*Andrews*, 1984; *Pitlick and Cress*, 2002; *Torizzo and Pitlick*, 2004; *Mueller et al.*, 2005; *Parker et al.*, 2007). A caveat for all of our results, however, is that caution should be exercised when considering reported numerical values due to the difficulty in independently determining the threshold of particle motion. We emphasize that this remains one of the most critical problems in determining coarse-grained sediment mechanics in natural rivers.



Figure 3.1: (a) DEM of North East Puerto Rico (inset) with Mameyes watershed outlined in red. The red and green circles represent the approximate locations of tracer particles in the main channel and headwaters stream (Bisley 3), respectively. Blue diamonds represent the location of USGS and USFS stream gaging stations, and the blue line represents the main channel of the Mameyes River, flow is from South to North. The red bounding rectangle represents the area in part (e). (b) Close up map of the headwaters stream showing the location of the tracer particles (green circles) at the time of the final survey. (c) Grain size distributions for the main channel site determined by Wolman pebble count for the channel (black line), initial population of tracers (red line), and second population of tracers (blue dashed line). (d) Grain size distributions for the headwaters site determined by Wolman pebble count for the channel (black line), and population of tracers (green line). (e) Close up map of the main channel field site showing the locations of the first (red circles) and second (blue circles) populations of tracer particles at the time of the final survey.



Figure 3.2: (a) Hydrograph for the duration of the study in depth (m) for the main channel of the Mameyes River. The dashed red lines represent two determinations of the critical shear velocity (m/s). (b) Hydrograph for the duration of the study in depth (m) for the headwaters field site of the Mameyes River for the duration of the study. The dashed red line represents the empirically determined critical shear velocity (m/s). Gray lines represent missing data.



Figure 3.3: (a) Photograph of the main channel of the Mameyes River looking upstream to the location where the tracer particles were installed. The width of the wetted portion of the channel is approximately 20 m. (b) Longitudinal profile extracted from a lidar DEM of the main channel of the Mameyes River with 0 m downstream representing the initial location of the tracer particles and 1200 m representing the approximate location of the final tracer found. (c) Photograph of the Bisley 3 stream looking upstream showing the location of the farthest tracer found downstream. The wetted region in the foreground is approximately 2 m wide. (d) Longitudinal profile from field survey, gray crosses represent the location of survey points. Tracers were installed 6 m downstream of the starting point, and the final tracer was located 10 m downstream of the last survey point.



Figure 3.4: Calculation of the dimensionless impulse (I_*) for two estimates of U_{*c} for a single flood from the main channel of the Mameyes River, where the time represents the floods location on the hydrograph in Figure 2a. The limits of integration for U_{*c1} and U_{*c2} are t_1 to t_4 , and t_2 to t_3 , respectively. The shaded region represents the region integrated for the calculation of I_* using U_{*c1} .



Figure 3.5: Fraction of mobile tracers (f) for single floods against peak Shields stress $(/tau_*)$. The red line represents the best fitting linear relationship, for which the intercept represents the critical Shields stress. See text for discussion of error bars.



Figure 3.6: (a) Frequency magnitude distribution of shear velocity for the main channel of the Mameyes river. The dashed red line represents an exponential function fit to the distribution for $U_* > U_{*c}$. (a. inset) Frequency magnitude distribution of discharge. The red line represents a power-law relationship, and the vertical dashed black line is the location of the threshold of motion. (b) Frequency magnitude distribution of the dimensionless impulse (I_*) . (c. inset) PDF of $ln(I_*)$, where $\langle U_* \rangle$ represents the average value of the distribution of $U_* > U_{*c1}$, $\langle t \rangle$ represents the average duration of a flood above the threshold of motion for the duration of the study, and D_{50} represents the median grain size of the tracer particles.



Figure 3.7: (a) Dimensionless displacement distributions for individual floods normalized by the mean $(\langle X/D \rangle)$ displacement for that flood. The black dashed line represents an exponential distribution. Dimensionless mean displacement lengths for each flood are labeled in the legend. (b) Contour density plot of X/D against the excess shear velocity normalized by the settling velocity for each tracer. The contour colors represent the density of tracers within that location. The value of U_* represents the flood peak, while the value of U_{*c} is treated as a fitting parameter. The black line represents the expected linear relationship between the dimensionless shear velocity and the modal tracer step length (Equation 3.1).



Figure 3.8: (a) Cumulative dimensionless displacement for each tracer survey for the first population of tracers. Survey number is denoted in the legend. (a. inset) Cumulative dimensionless displacement for the second population of tracer particles, installed immediately prior to survey 4. (b) Cumulative dimensionless displacement data normalized by the mean displacement for each survey for tracer population one. (b. inset) Cumulative dimensionless displacement for each survey for tracer population one.



Figure 3.9: Mean displacement data for the first (red +) and second (blue x) populations of tracer particles vs dimensionless impulse (I_*) . The gray line is the linear relationship determined by *Phillips et al.* (2013). The black lines represent the upper and lower limits of sediment particle transport. Limit 1 represents a single step per transporting event, while limit 2 represents a tracer experiencing continuous motion for the duration of a flood. The data plotting below Limit 1 have unrealistic values for I_* , which are due to small values of I_* in this region being particularly susceptible to the determination of U_{*c} (see Section 3.3 for explanation).



Figure 3.10: Mean displacement data for the first (red +) and second (blue x) Mameyes tracers, and Bisley 3 tracers (green o). The solid and dashed black lines represent a linear relation between $\langle X/D \rangle$ and I_* for the Mameyes tracers, and the Bisley 3 tracers, respectively. I_* is calculated using U_{*c2} .



Figure 3.11: (a) CDFs for tracer grain size (mm) for single floods separated by whether the tracer moved (light gray lines) or remained immobile (black lines). The red and blue lines represent lognormal functions fit to the combined mobile and immobile tracers, respectively. (b) Tracer grain size normalized by D_{50} against tracer travel distance normalized by the expected step length (X_s) from equation 1. Red crosses represent mobile tracers for all single floods near the threshold of motion, blue diamonds represent tracers that did not experience movement (n=269) (solid diamonds represent multiple tracers plotted on top of each other). The dashed black line denotes the expected single step length.


Figure 3.12: (a) Tracer grain size (mm) against distance (km) the particle has traveled for the first (red +) and second (blue x) tracer populations at the final survey. Black and gray squares represent the grain size percentiles at the corresponding distance downstream for the river. The dashed lines are moving averages to guide the eye. (b) Spatial standard deviation for the river, and both populations of tracer particles against dimensionless distance (X_*). The lines represent fitted exponential functions with the form of equation (3.3). (c) Coefficient of variation moving downstream for the river and both populations of tracer particles. (d) Collapse of the stream and tracer populations standard deviation data. σ_o represents the coefficients from the fitted equations in (b). (e) Downstream decrease in the mean (\overline{D}) grain size. Lines represent equation (3.4).



Figure 3.13: Collapse of the tracer data for the first (red +) and second (blue x) Mameyes tracer populations, and Bisley 3 tracers (green o). The black line represents a linear relationship between the mean particle displacement and the dimensionless impulse over the dimensionless friction factor (equation 3.5).

CHAPTER 4 : The threshold of motion as a filter on extreme climate events

Abstract:

Bed load transport is a key regulator of landscape denudation and channel form, through its control on (1) erosion rates of bedrock streams and (2) channel geometry of alluvial rivers. A useful simplifying principle in assessing long-term rates of sediment transport is that the wide range of flow events generated by climate may be distilled to a single characteristic, 'bankfull' flood. This concept has led to the successful formulation of equilibrium channel theory for gravel rivers, in which fluid stress slightly exceeds the threshold of motion during bankfull conditions. Work over the last decade has challenged this paradigm, however; in particular, the observation that many rivers exhibit a power-law distribution of flood magnitudes has been interpreted to suggest that (1) a characteristic flood, and hence equilibrium, may not exist, and (2) that extreme events dominate landscape erosion rates. Here we demonstrate that the threshold of motion represents a first-order filter on the statistical scaling of floods in gravel-bedded alluvial and bedrock rivers. Results are drawn from a detailed study of a single river, and from a compilation of stream gage data from 197 locations across the United States. While many of these rivers indeed exhibit a power-law distribution of discharge magnitudes, the distribution of fluid stress for flows above the threshold of motion is exponential for nearly all cases. Moreover, data reveal that gravel and bedrock rivers both have geometries adjusted to near-threshold bed load transport. Extreme events in discharge may be common, but the coupled evolution of transport and channel geometry dictates that extreme events in fluid stress are exceedingly rare. These findings reaffirm and unite two classic principles of geomorphology - equilibrium channel theory and the bankfull flood - and cast doubt on the importance of extreme events as dominant mechanisms of river evolution.

4.1. Introduction

Through the erosion and deposition of sediment, rivers represent the mechanism through which climatic shifts in precipitation and the rise and fall of the oceans are propagated through landscapes (e.g., *Tucker*, 2004). Rivers, in conjunction with glaciers, erode mountain ranges and transport the resulting sediment to the sea. Bed load transport, through its control on bed rock erosion (Sklar and Dietrich, 2004), may be the limiting factor in rates of mountain landscape incision (e.g., Snyder et al., 2003; Molnar, 2004). Gravel rivers exist throughout many climatic regimes and experience drastically different climatic forcing, yet they are universally observed to be adjusted to near threshold conditions (Leopold and Maddock, 1953; Parker et al., 2007). Even bedrock channels may be adjusted to near-threshold bed load transport (*Phillips and Jerolmack*, 2014), though this has yet to be demonstrated broadly. Diverse climatic regimes, in concert with the physical properties of a drainage basin, produce a diverse population of floods. A key guiding principle in geomorphology is that the wide range of fluid stress magnitudes and recurrence intervals may be reduced to a single dominant discharge that shapes the landscape; it emerges as the peak in the distribution of the product of stress magnitude and recurrence frequency (Wolman and Miller, 1960). This peak represents the flood whose combination of frequency of occurrence and magnitude of transport moves the most sediment in a time integrated sense; it is rarely the most frequent small floods or the infrequent catastrophic floods (Wolman and Miller, 1960). This flood is commonly referred to as the 'bankfull' flood, as it represents the flood for which the channel is adjusted (Wolman and Miller, 1960; Wolman and Gerson, 1978). Indeed, equilibrium channel theory is predicated on the existence of a characteristic flood. For coarse-grained rivers, channel geometry adjusts to the channel-forming discharge to achieve an equilibrium, in which fluid shear stress (τ) is slightly in excess of the threshold of motion (τ_c) , $\tau = (1 + e)\tau_c$ (Parker, 1978, 1979; Parker et al., 2007). Field observations, laboratory experiments, and analytical theory show e = 0.2 - 0.6, yielding $\tau/\tau_c = 1.2 - 1.6$ at the bankfull depth (Parker, 1978, 1979; Pitlick and Cress, 2002; Torizzo and Pitlick, 2004; Mueller et al., 2005; Parker et al., 2007; Pitlick et al., 2013). Despite the complexity of sediment transport, the crystallizing simplicity of the equilibrium threshold channel suggests that the complexities of climatic forcing may be greatly reduced for geological timescales. In particular, the hydrograph may be represented with a single value for fluid stress and an intermittency factor I = t/T, where t is the duration of time that a flow exceeds τ_c and T is the total duration of interest. While this approximation is seemingly an oversimplification, it has been shown to adequately describe the hydrology over annual timescales in a recent study of bed load tracer transport (*Phillips et al.*, 2013). This formulation should hold as long as the frequency-magnitude distribution of the hydrograph remains thin-tailed, and the observation window is long enough for this distribution to converge (*Phillips et al.*, 2013). Equilibrium channel theory implies that, for a channel adjusted to a particular climatic regime, the distribution of flood stress is restricted to values that do not exceed the threshold of motion by too large a value.

However, the statistical scaling of discharge magnitudes is a matter of continued debate, with models falling roughly into two classes; those implementing exponential tails (e.g., Tucker and Bras, 2000; Snyder et al., 2003; Tucker, 2004), and those using power-law tails (e.g., Laque et al., 2005; DiBiase and Whipple, 2011; Laque, 2014). It has been noted that the choice of tail distribution can result in significantly different relations between channel steepness and erosion rates (*DiBiase and Whipple*, 2011). An exponential tail supports the notion of near-threshold equilibrium channel geometry, while power-law scaling largely ignores this question by assuming that threshold constraints can be neglected as floods of interest greatly exceed the threshold (Baldwin et al., 2003). There are compelling reasons to believe that discharge follows a power-law distribution over some scaling range (Turcotte and Greene, 1993; Turcotte, 1994; Malamud and Turcotte, 2006). This observation, coupled with a non-linear dependence of river incision on stress (e.g., Howard and Kerby, 1983; Howard, 1994), has led to the assertion that the importance of extreme events as agents of landscape evolution has been overlooked (Molnar, 2001, 2004; Molnar et al., 2006). More specifically, researchers have argued that the power-law scaling of discharge indicates that climate variability exerts a primary control on net landscape denudation (*Peizhen* *et al.*, 2001; *Molnar*, 2001, 2004; *Lague et al.*, 2005; *Molnar et al.*, 2006). If true, the dominance of rare, extreme events would also suggest that many river channels never achieve an equilibrium channel geometry, because there is no characteristic flood. Thus we are presented with something of a paradox on the importance of extreme versus near-threshold hydrology as dominant mechanisms of landscape evolution.

Here we present the results of a case study for a river in Puerto Rico that experiences extreme fluctuations in discharge due to frequent, short-duration high-magnitude precipitation events. We show that, although discharge scaling approaches a heavy-tailed power-law, the channel geometry is adjusted to near-threshold conditions throughout the catchment. We also show that the distribution of fluid stresses for flows in excess of the threshold of motion is exponential, not power law. There is not only a characteristic stress associated with the channel form, but there is a characteristic flood duration, such that that the channel is adjusted to a characteristic impulse (*Phillips et al.*, 2013) - flow magnitude times duration. Furthermore, we demonstrate through a compilation of stream flow records from several climatic regimes across the United States, that the results from this case study are the rule, rather than an exception.

4.2. Case study: the Mameyes River

The Mameyes River is nestled in the heart of the Luquillo Mountains and the Luquillo Critical Zone Observatory (Figure 4.1(a)), located on the Northeastern corner of Puerto Rico. Due to a strong orographic effect the Luquillo Mountains receive, on average, over 4000 mm/yr of precipitation. Frequent convective storms, including tropical depressions and hurricanes, coupled with steep topography result in regular flash flooding (Figure 4.1(b-c)) capable of mobilizing coarse sediment (*Schellekens et al.*, 2004; *Pike et al.*, 2010; *Phillips et al.*, 2013). The Mameyes River is characterized by extreme fluctuations in stream flow (recorded by the United States Geological Survey (USGS) stream gage 50065500, drainage area of 17.8 km²), with discharges (Q, m³/s) over the last 20 years commonly exceeding 300-400 times the base flow value (Figure 4.1(b)). However, base flow does not erode

coarse sediment, which is required to further incise the landscape. Previous work (*Pike*, 2008; Phillips et al., 2013; Phillips and Jerolmack, 2014) has determined that the critical discharge (Q_c) to entrain coarse sediment $Q_c=6.6 \text{ m}^3/\text{s}$, indicating that the river is only transporting sediment 3.5% of the time (Figure 4.1(b)). Despite the mostly quiescent nature of the Mameyes River, it remains highly variable with peak discharges reaching values of up to 65 times Q_c . This large range in discharge is evident in the heavy-tailed powerlaw scaling of the frequency-magnitude distribution of discharge (Figure 4.1(c)). However, despite the extreme range in forcing, *Phillips et al.* (2013) showed through the frequent tracking of RFID-transponder tagged cobbles and hydrological analysis that the sediment transport and hydrology in alluvial and bedrock (Figure 4.1(a)) sections remains adjusted to near-threshold conditions. Furthermore, *Phillips and Jerolmack* (2014) observed that, for bed load transporting events, there was a rapidly diminishing probability of occurrence for flows exceeding threshold. An important point is that the scaling of discharge is misleading; flow depth (h) is proportional to fluid stress, which is the relevant parameter for sediment transport, but depth is a slowly-varying function of discharge (for the Mameyes, $h \sim Q^{0.3})$ (Leopold and Maddock, 1953; Leopold et al., 1964; Parker et al., 2007). For flows exceeding bankfull, the relation between depth and discharge often becomes even weaker (Dodov and Foufoula-Georgiou, 2005).

The tracer cobble results indicate near-threshold, partial bed load transport for the Mameyes River for the duration of the tracer study (2010-2012). However, the tracer results do not sample the largest events and therefore one may remain skeptical of extrapolating these results to longer timescales. However, the channel geometry integrates over innumerable flooding events, providing an opportunity to test whether the channel is adjusted to nearthreshold conditions at longer timescales. To account for changes in slope and particle size downstream we use the Shields stress (τ_*),

$$\tau_* = \tau_b / (\rho_s - \rho) g D \tag{4.1}$$

a nondimensional shear stress, where τ_b is the basal shear stress (Pa), ρ_s is the density of sediment (2650 kg/m³), ρ is the fluid density (1000 kg/m³), g is the acceleration due to gravity (m/s^2) , and D is the median particle diameter (m). It is common to use the normal flow approximation to provide an estimate of the boundary shear stress, such that $\tau_b = \rho g h S$, where S is the channel slope. To test whether the Mameyes is adjusted to near-threshold conditions we compute the ratio of the bankfull and critical Shields stresses (τ_{*bf}/τ_{*c}) along the channel from a small steep step-pool headwaters channel to the gravels and transition. To compute τ_{*bf} we extract the longitudinal profile and slope of the channel from aerial lidar data (Figure 4.2(a-b)), and compute the particle size from random walk pebble counts (Wolman, 1954) spaced down the channel (Figure 4.2(c)). We reanalyze channel cross sectional data collected by Pike (2008) to determine the relationship between bankfull discharge Q_{bf} , and the average channel depth at bankfull, $\langle h_{bf} \rangle$ (Figure 4.2(d)), where $\langle \rangle$ denotes the arithmetic average. We use the close agreement between Q_{bf} and $\langle h_{bf} \rangle$, and Q_{bf} and downstream distance (Figure 4.2(e)) to compute how $\langle h_{bf} \rangle$ changes downstream. Measurements of τ_{*c} are rare in general, and our previous tracer work was able to estimate it for only two locations within the Mameyes watershed. It is common to assume a constant value of τ_{*c} , but τ_{*c} has been observed to vary both spatially and temporally, and possesses a strong correlation with channel slope (Mueller et al., 2005; Lamb et al., 2008). Therefore we compute τ_{*c} from an empirical relation (Mueller et al., 2005) with channel slope

$$\tau_{*c} = 2.18S + 0.02. \tag{4.2}$$

Although we have few measurements, we find good agreement between values for threshold predicted from equation (4.2) and the two empirical estimates of the threshold from our previous tracer cobble study (*Phillips and Jerolmack*, 2014). Moreover, the measured values bracket the range of slopes within the stream lending support to the applicability of equation (4.2) to the Mameyes catchment, and thus we apply equation (4.2) with no calibration. There is growing suspicion that τ_{*c} is a dynamic variable that can change between floods, and within a flood (*Hsu et al.*, 2011; *Turowski et al.*, 2011; *Marquis and Roy*, 2012), however we currently lack theoretical, analytical, or empirical means to account for these effects. Although there is considerable scatter, we find that the ratio of τ_{*bf}/τ_{*c} exhibits no downstream trend, and fluctuations around a constant value $\langle \tau_{*bf}/\tau_{*c}\rangle = 1.7$. Despite the expectation from RFID tracer results that the Mameyes River is a adjusted to near threshold conditions, it is remarkable that the channel geometry confirms this expectation for over three orders of magnitude in channel slope and across vastly different channel morphologies. Thus despite exceptional floods, including two category 3 hurricanes, the channel geometry reflects near threshold conditions. Importantly, the upper portions of the Bisley watershed (where slopes exceed 10%) are alluviated bedrock channels; despite the presumed control of bedrock on channel slope, headwater streams have nonetheless adjusted their geometry to near-threshold transport conditions.

In light of this threshold scaling, we focus on the statistical scaling of the hydrograph (20 years) for the Mameyes River. Here we utilize the shear velocity (m/s)

$$U_* = \sqrt{\tau/\rho} = \sqrt{ghS} \tag{4.3}$$

to discuss the statistical scaling of the hydrograph, because it has been demonstrated that bed load displacement is proportional to shear velocity (*Lajeunesse et al.*, 2010; *Martin et al.*, 2012; *Phillips and Jerolmack*, 2014). The frequency-magnitude distribution of U_* (m/s) possesses power-law scaling, though we note that the transformation from Q to U_* results in a significantly steeper power-law decay (Figure 4.3(a)). The frequency-magnitude scaling of U_* (m/s) has an upper truncation at approximately the threshold of bed load motion, beyond which the distribution is well described by an exponential (Figure 4.3(a)). That the scaling is well fit by an exponential distribution $(p(U_* > U_{*c}) = Ce^{-\lambda})$ is significant, in that an exponential function is completely described by its mean value, λ^{-1} . The mean value $\lambda^{-1} = (\langle U_* \rangle_{U_*>U_{*c}} - U_{*c})$; for the Mameyes, we find then that the mean shear velocity for flows above the threshold of motion is $\langle U_* \rangle_{U_*>U_{*c}} = 0.38$, which is in close agreement with the independently determined $U_{*bf} = 0.402$ from the morphologic survey of the channel ($Pike \ et \ al., 2010$).

Despite the close agreement between $\langle U_* \rangle_{U_* > U_{*c}}$ and U_{*bf} , a flood contains a range of values for U_* throughout the rise and fall of the hydrograph. This variability has previously (*Phillips et al.*, 2013; *Phillips and Jerolmack*, 2014) been shown to be approximated by the dimensionless impulse:

$$I_* = \int_{t_s}^{t_f} (U_* - U_{*c}) \,\mathrm{d}t / D_{50}, \quad U_* > U_{*c}$$
(4.4)

which represents the integral of the excess fluid-derived momentum imparted to the channel bed, where t_s and t_f are the starting and finishing times of the flow of interest. The probability density function (PDF) of I_* (Figure 4.3(b)) is at first inspection a negativelyskewed lognormal distribution, with a primary mode centered at ~400-1000 $(ln(I_*)=6-$ 7). The primary mode of the PDF of I_* coincides with an independent calculation of the bankfull impulse $I_{*bf} = (\langle U_* \rangle_{U_* > U_{*c}} - U_{*c}) \langle t \rangle / D_{50}$ where $\langle t \rangle$ represents the average duration of the distribution of flood durations above the threshold of motion (~ 1.4 hrs for the Mameyes), and we reiterate that $\langle U_* \rangle_{U_* > U_{*c}}$ represents the average of the entire record for $U_* > U_{*c}$. This strongly indicates that not only is there a channel forming flow, but that channels are adjusted to a characteristic impulse; this is a demonstration of the principle of maximum geomorphic work (Wolman and Miller, 1960), in that the peak in the impulse distribution represents the flood whose combination of frequency of occurrence and magnitude in shear velocity is most responsible for the transport of sediment over the long term. On closer inspection of the hydrograph we see many small values of I_* (less than ~ 10 ; these represent artifacts of the resolution of a non-continuous time series and the application of a depth-averaged stress (for a discussion of this problem see *Phillips and* Jerolmack (2014)). Coupling these reservations, with the caveat that computation of I_* is also exceptionally sensitive to the value of U_{*c} (Phillips and Jerolmack, 2014), we are comfortable stating that the higher frequency of the smallest values is a product of the analysis and does not likely represent real phenomena.

The Mameyes River is adjusted to near-threshold conditions along its entire length, despite being partially controlled by bedrock in some locations, and experiencing exceptionally large precipitation events. We hypothesize that, because the Mameyes River is an end member in terms of discharge variability and channel slopes, that thin-tailed scaling of sedimentmoving fluid stresses and the existence of a characteristic bankfull flood are results that extend to all gravel-bedded alluvial and bedrock channels.

4.3. The statistical scaling of coarse-grained rivers

In this section we test the generality of the results from the Mameyes River with a large data compilation of gravel rivers located at USGS stream gaging stations across the United States.

4.3.1. Methodology

Stream flow gaging stations operated by the USGS are located across the United States, however there are a surprisingly small number of field sites that have the requisite data needed to test the generality of the Mameyes River case study. In order to use a field site it must meet the following five criteria: (1) reported measurements of channel bed slope, (2) particle size (D_{50} ; 10 mm), and (3) channel hydraulic geometry, (4) be located near a USGS stream gage with greater than 10 years of instantaneous discharge records, and (5) provide an accurate stage-discharge rating curve. Out of over 1000 field sites examined, 197 met the five requirements above (Figure 4.4). These sites have the following ranges for relevant variables: $0.0002 \le S \le 0.14$, $11 \le D_{50} \le 213$ mm, $2.62 \le$ drainage area ≤ 30574.7 km², and $\sim 800 \le$ precipitation ≤ 4000 mm/yr. Precipitation data represent 30 year normals from the PRISM Climate Group (Oregon State University, http://prism.oregonstate.edu, created 2013).

The processing steps undertaken for each stream gage are illustrated in Figure 5 where we use USGS gage 03151400 (Little Kanawha River nr. Wildcat, WV) as an example. Stream flow time series data are from the USGS Instantaneous Data Archive (IDA, before

2007) (Figure 4.5(a)) and the USGS National Water Information System (NWIS, after 2007) (Figure 4.5(b)). Channel slope, particle size and bankfull depth data are derived from peer-reviewed scientific articles, USGS reports, and US Forest Service reports. We only use field sites with slopes determined from field surveys of the channel bed or water surface, or from aerial lidar measurements. For consistency we only use particle size data determined from random walk pebble counts, and we do not use grain size data classified by the Wentworth scale. We consider the reach a gravel stream for the purposes of this analysis if the D_{50} is greater than 10 mm. At a large majority of the selected field sites Q_{bf} and $\langle h_{bf} \rangle$ were surveyed. The majority of streams are alluvial, however a fair number possess large bedrock outcrops, are bedrock influenced, or are thinly alluviated bedrock rivers. Due to the numerous methodologies from which the bankfull depth is measured in the field (Williams, 1978), we only use estimates of bankfull depth that were determined using compatible methodologies. For the large majority of gages used the bankfull depth represents the average depth of the channel at the height of the floodplain. In cases where there was no broad floodplain, the bankfull depth was recorded as a significant break in slope on the channel bank. In all cases we checked the field surveyed bankfull depth against the USGS rating curve. We have only used measurements of bankfull depth that are related to a morphologic break, excluding those that are solely determined from vegetation, flood recurrence, and other non-morphologic measures. Measurements of the channel geometry are available at all USGS stream gages as these variables are surveyed repeatedly to establish the rating curve between stage (flow depth relative to the gage) and discharge. The USGS uses a variety of methods to measure discharge during surveys, depending whether the flow in question is wadeable or needs to be measured from a suspended cableway or nearby bridge. Only discharge (not stage) is archived as an available long term data set, therefore we compute the time series of channel stage using the USGS rating curve (Figure 4.5(c)). Unfortunately, at many field sites the values of stage do not represent the actual flow depth, and thus result in unphysical values of U_* . Thus we turn to the USGS field measurements (Figure 4.5(d)) used to create the rating curve to determine a reliable depth (Figure 4.5(e)).

For those flows at which the stream is wadeable, the width (W, m), cross sectional area $(A_{xs},$ m^2), stage (h_s , m) and flow velocity are recorded. The cross sectional area is computed from the width and $\langle h \rangle$, and thus we acquire $\langle h \rangle$ from recorded values of A_{xs}/W . However, we cannot simply compute a relationship between the measured depths and their corresponding values of discharge, because depth is not measured during high flows. We also cannot extrapolate from the measured depth data, because it is well documented that many depth discharge relationships have several separate scaling regions with a prominent change in scaling typically occurring at the top of the banks (Dodov and Foufoula-Georgiou, 2005). To circumvent this problem we determined a relationship between h_s and $\langle h \rangle$; we found that the relation is typically of the form $\langle h \rangle = h_s - K$, where K is a constant specific to each stream. Thus, we fit a linear function to log-binned values of h_s and $\langle h \rangle$ (Figure 4.5(f)), and this linear relation is then used to transform the stage-discharge rating curve into a $\langle h \rangle$ -discharge rating curve (Figure 4.5(g)). We can then use equations (4.1) and (4.3) to compute time series of τ_* and U_* . We are able to test the validity of this approach using the independently surveyed values of the bankfull depth $(h_{bf-survey})$ and discharge $(Q_{bf-survey})$, because $Q_{bf-survey} = Q_{bf-shifted} = Q_{bf-unshifted}$. We find generally that, for a given value of $Q_{bf-survey}$, the corresponding values $h_{bf-survey} \approx h_{bf-shifted} \neq h_{bf-unshifted}$ (Figure 4.6(a)). This processing (Figure 4.5(a-g)) ultimately results in an objective and reproducible method that shifts the frequency-magnitude distribution of fluid stress. As an example, we show in Figure 4.6(b) that using h_s to calculate the shear velocity results in U_* at baseflow exceeding U_{*bf} , which is a non-physical result for this coarse-grained channel ($D_{50} = 91$ mm). We note that not all records require a shift, and we use in the following analysis the rating curve that best represents the $\langle h \rangle$ data.

A common problem with long term data records is that the data do not often represent a single sampling rate. We find that the inclusion or exclusion of these data does not affect the observed results, but it can introduce numerical artifacts when computing the integral in equation (4.4). Therefore, when calculating I_* we resample the U_* time series with a frequency of 1 Hz by linearly interpolating between the sampling points. This allows us to eliminate numerical artifacts at the edges of each flood caused by the coarser resolution dataset. For the following analysis we use equation (4.2) to determine an estimate of the threshold of motion, except for 13 gages where the threshold of motion was empirically determined from measurements of bed load transport (see *Mueller et al.*, 2005). We exclude gages for which equation (4.2) predicts a threshold that results in the river being above the threshold of motion 30% of the duration of the entire record, and gages for which U_{*c} excludes 99.9% of the record. These criteria serve as further quality control, as we are lacking a rigorous method for which to determine the threshold of motion with the data that are widely available at most streams. The second criterion excludes the records with insufficient data to determine a reliable statistical distribution of I_* . We note that these two criteria result in the exclusion of 43 gages.

4.3.2. Frequency-magnitude and dimensionless impulse scaling

We compute the frequency magnitude distribution of U_* for 153 stream gages. We find that for all stream gage records examined that the distribution of $U_* > U_{*c}$ is thin-tailed, and in the majority of cases well described by an exponential function. We expect that a short hydrological record is the cause for the small number of distributions of U_* that decay faster than the exponential distribution. There is a wide range of scatter among the many stream gages, though this may be expected as the range of stream gages represents several different climatic regimes. That the records of $U_* > U_{*c}$ are well described by an exponential distribution with a single fitting parameter gives us reason to believe that a normalization by $\langle U_* \rangle_{U_*>U_{*c}}$ could produce a collapse of the data. Thus we normalize each frequency-magnitude distribution by its average value $\langle U_* \rangle_{U_*>U_{*c}}$. We find that this produces a reasonable collapse of the data (Figure 4.7) for which the mean scaling is very well described by an exponential function for flows above the normalized threshold of motion (Figure 4.7). We observe that the tail decays faster than the exponential function for the largest 0.15% of $U_* > U_{*c}$, but this is to be expected due to under sampling. We might expect that a longer stream flow record of $U_* > U_{*c}$ would converge to an exponential function for the tail. Here we note that $\langle U_* \rangle_{U_* > U_{*c}} / U_{*c} = 1.1$, which when converted to Shields stress gives a ratio of $\tau_* / \tau_{*c} = 1.23$; this is very close to the analytical prediction of $\tau_* / \tau_{*c} = 1.2$ (*Parker*, 1978, 1979) for alluvial gravel channels.

An intriguing implication of the Mameyes results is that the mean of the exponential function is approximately the bankfull flood. To test the generality of this result we compare $\langle U_* \rangle_{U_*>U_{*c}}$ to the value U_{*bf} determined using surveyed bankfull depths near each stream gage (Figure 8). We find remarkably good agreement between the two, in that $\langle U_* \rangle_{U_*>U_{*c}} \approx U_{*bf}$ for a large range of channel slope (Figure 4.8(inset)). The agreement is best when using field sites for which the threshold of motion has been empirically determined, suggesting that much of the scatter in Figure 4.8 may be due to errors in estimating U_{*c} from equation (4.2).

Lastly, we compute the distribution of I_* for 153 stream flow gages. We find that the large majority of gages are represented by negatively-skewed unimodal log-normal distributions. However, there are numerous individual gages which display a secondary mode for very small values of I_* , which potentially gives rise to the negative skew. Inspection of the hydrograph for gages with prominent secondary modes shows that these stream flow records possess short sections for which the hydrograph fluctuates around the threshold of motion, effectively producing a series of single-point floods. It is likely that these values are errors, as we can find no mention within USGS records of flow control structures upstream of the gages in question that could produce such short-duration, pulsed floods. Therefore we filter the impulse data to include only those floods with durations lasting longer than 1 hr (typically 1-4 sampling points), which excludes the majority of the error prone sections of the hydrographs. Based on the scaling of I_* within the Mameyes River, we normalize the respective impulse distribution of each stream by the product of the average excess shear velocity and average flood duration, $(\langle U_* \rangle_{U_* > U_{*c}} - U_{*c}) \langle t \rangle / D_{50}$. We then calculate the histogram of the combined I_* , normalized I_* data from all streams, and find that it is well represented by a negatively-skewed log-normal distribution (Figure 4.9) centered on

one (zero in natural-log space). This strongly suggests that the channel is adjusted to a characteristic flood impulse. Further, the data imply that extreme events, in terms of flood magnitude and duration, are unlikely to occur in gravel-bedded streams. We note here that the distribution would likely contain more values on the positive axis if the hydrological record were complete, however it is unlikely that this distribution would develop a heavytail.

4.4. Discussion

The results from the Mameyes River case study show that the threshold of coarse sediment transport exerts a first order control on the distribution of applied stress. We posit that this should remain true for any channel in which bed load transport is the driving process shaping the channel – including both alluvial rivers and alluviated bedrock channels. The Mameyes River exemplifies this point in that the distribution of Q approaches that of a heavy-tailed distribution (Figure 4.1(c)), for which the variance would not converge. Despite this, the distribution of $U_* > U_{*c}$ for the same time period is well described as an exponential distribution (Figure 4.3(a)). In light of the generality of the Mameyes River results (Figure 4.7), this case study shows that the existence of power-law scaling of floods does not lead to power-law scaling of $\langle U_* \rangle_{U_*>U_{*c}}$. In other words, a heavy-tailed distribution of Q is not evidence for the importance of extreme events in the evolution of landscapes. This is because the threshold of motion in bed load channels acts as a filter on extreme events in precipitation; the rivers ability to adjust its width and depth to near-threshold conditions acts to decouple the distribution of U_* from the distribution of Q imposed by the climate.

That the distribution of $\langle U_* \rangle_{U_*>U_{*c}}$ is exponential is non-trivial, as this distribution suggests that the probability of observing a flow with applied stresses greater than the threshold of motion diminishes rapidly with increasing magnitude. To the authors knowledge, this is the first demonstration that physically links a statistical distribution of flow events to sediment transport mechanics, and channel form; the bankfull flood that shapes the channel represents the mean value of excess shear velocity (or stress). An important implication is that, with only measurements of the average bankfull depth and channel slope, one can determine an average threshold of motion as $U_{*c} = U_{*bf}/1.1$. The value of 1.1 is for the average U_{*c} in Figure 4.7, with the addition of a sufficiently long hydrograph one can calculate U_{*c} numerically for each site. Another more speculative but attractive idea suggested by the collapse of data in Figure 4.7 is that, given an estimate for U_{*bf} , one could invert the exponential distribution to predict flood statistics.

This exponential scaling might also explain why we observe channels adjusted to a range of values of U_*/U_{*c} larger and smaller than the analytical value of 1.1 ($\tau_*/\tau_{*c} = 1.2$). In averaging over innumerable floods the channel might approach the analytical prediction, as is the case in Figure 4.7, but because the distribution of forcing events is stochastic the channel is randomly sampling this exponential tail. Thus, the channel may adjust its geometry to some unknown number of the floods that it has recently experienced. Even an exponential distribution predicts that large values of U_* will occasionally occur; we might expect that these flows are generally destructive to the channel geometry resulting in a substantially wider channel (*Wolman and Gerson*, 1978). However, because the probability of these events is small, their contribution to time-integrated bed load transport is likely limited, and their signature on channel geometry is erased by smaller more frequent floods (*Wolman and Gerson*, 1978). To understand the link between channel geometry at one location and the floods that were responsible for that shape, one would need detailed knowledge of the timescales of channel adjustment, which is largely unknown (*Wolman and Gerson*, 1978).

The timescales over which rivers adjust to perturbations are not particularly well quantified; however the limited observations available suggest that they are fairly rapid when compared to timescales of climatic change (*Wolman and Gerson*, 1978; *Johnson et al.*, 2010). Also, not all changes in forcing or tectonics lead to enhanced incision and erosion as a channels ability to adjust its width has been shown to be a first order control on channel response to perturbations (*Yanites et al.*, 2010). Furthermore detailed observations of incision due to extreme events and moderate events have shown that only the moderate events are capable of producing the long term incision rate (*Hartshorn et al.*, 2002; *Barbour et al.*, 2009). We do not have many observations of the impact of climate change on channel geometry, but recent work by *Slater and Singer* (2013) has shown that the adjustment of channels to climatic change is observable in available hydrologic records (1950 to 2012), indicating a rapid response. However, there are numerous observations of other processes that affect catchment hydrology, such as urbanization (e.g., *Chin*, 2006). For example, increased peak flows (*Leopold*, 1968) due to the application of impervious surfaces result in a rapid (years to decades) increase in channel width followed by slow relaxation due to decreased sediment loads (*Wolman*, 1967). Thus we find it highly unlikely that that climatic variability might lead to a regime in which rivers are unable to reach an equilibrium condition, a concept that has been suggested by some researchers to be responsible for a net enhancement in landscape erosion (*Peizhen et al.*, 2001; *Molnar*, 2004).

The statistical scaling of I_* for all examined rivers strongly reinforces the concept of the near-threshold channel adjusted to a dominant flood. Not only are rivers adjusted to a characteristic stress, but there is also a characteristic flood duration; the product of these two produce a characteristic impulse for a river. Lacking a strong theoretical framework for how flood durations and I_* should scale, we are hesitant to rule out the negatively skewed part of the distribution. We might expect, however, that with a complete record the distribution would shift positively such that the peak is centered at the bankfull flood. The thin-tailed behavior of $U_* > U_{*c}$ indicates that the integral in equation (4.4) can be replaced by a single characteristic stress times an intermittency factor. Thus, when examining timescales significantly longer than the recurrence interval of a bed-load-transporting flood, equation (4.4) reduces to $\langle U_* - U_{*c}\rangle IT$, or $(1 + e)\tau_{*c}IT$, which is the hydrologic description employed in some models of fluvial landscape evolution over geologic time (*Paola et al.*, 1992a). While, this hydrologic description seemed too simplistic at first glance, these results suggest that perhaps it is just simple enough.

4.5. Conclusion

We have shown through a case study in a flashy stream and a compilation of stream flow data that the threshold of motion for bed load transport exerts a first-order control on the stress applied to the river bed. We have also demonstrated that the existence of heavytailed power law statistics in discharge may result in larger flood volumes, but that it does not result in larger applied stresses due to the filtering effect of sediment transport. The threshold of particle motion, both a driver and consequence of evolved channel geometry, truncates the distribution of forcing such that the probability of events beyond this threshold diminishes rapidly. As stress, not discharge, is the relevant parameter for moving sediment; models that parameterize bed load transport or erosion as a function of discharge exaggerate the importance of extreme events. Remarkably, these results extend to many bedrock rivers as well, indicating that even channels with strong geological controls are adjusted to near-threshold transport during a characteristic flood. This work upholds and extends two central tenets of geomorphology; (1) equilibrium channel theory (e.g., Leopold et al., 1964; Parker, 1978) and (2) the principle of maximum geomorphic work (Wolman and Miller, 1960). The threshold filtering effect casts doubt on proposed linkages among extreme events, climate variability and landscape incision (*Peizhen et al.*, 2001; *Molnar*, 2001, 2004; Molnar et al., 2006). Our analysis also runs counter to claims that the full stochastic distribution of climate forcing – manifest through discharge – must be explicitly incorporated into landscape evolution models (see reviews by Tucker and Hancock, 2010; Laque, 2014). Instead, we find that all of the climatic variability encoded in the hydrologic record may be crystallized into a single parameter when considering bed load transport; the product of the bankfull shear velocity and an intermittency factor. This parameter has already been demonstrated to predict the downstream transport of tagged bed-load tracers (Phillips et al., 2013; Phillips and Jerolmack, 2014), providing a mechanistic link between hydrology and landscape evolution. That bed load transport is a key regulator of landscape erosion rates and channel form is well established (Sklar and Dietrich, 2004, 2008). The upshot of this work is that long term rates of bed load transport, channel evolution and

landscape erosion may be more predictable than we thought.



Figure 4.1: (a) Map of Mameyes River catchment, where green and red circles indicate locations of RFID tracer studies, blue diamond indicates location of USGS stream gage 50065500, and light blue line indicates section for which the longitudinal profile is derived. Inset shows location of the Mameyes catchment in Northeast Puerto Rico. (b) Hydrograph of 15 minute instantaneous discharge for the Mameyes River, with threshold of motion demarcated as a dashed red line. (c) Frequency-magnitude scaling of discharge for the hydrograph in (b).



Figure 4.2: (a) Longitudinal profile of the Mameyes River derived from aerial lidar measurements. (b) Channel slope where black + represent averages of channel slope in 200 m increments (10 times the channel width). Red lines represent best fit power functions. (c) Grain size measurements from random walk pebble counts. Red line indicates an exponential function, where dashed section indicates a constant value over the upper 1.5 km, as described in Litwin-Miller et al. (submitted). (d) Relationship between bankfull discharge and average depth at bankfull. The red line is a best fit power function. (e) Relationship between downstream distance and bankfull discharge, where red lines indicate a piece-wise power function. The transition does not vary smoothly due to the large difference in drainage area between the headwaters river and the main channel. (f) Measurements of τ_{*c} from RFID tracer cobbles show good agreement with equation (2). (g) Variation of the ratio τ_{*bf}/τ_{*c} with distance downstream. We note that there are no statistically significant trends in these data.



Figure 4.3: (a) Frequency-magnitude distribution of U_* (blue line) for the Mameyes River. Distribution of U_* possesses a truncation at approximately U_{*c} , beyond which the tail is well described by an exponential function, where the red diamond and blue circle represent U_{*bf} and $\langle U_* \rangle_{U_* > U_{*c}}$, respectively. The over fit for the upper 0.001% is likely due to under sampling. (b) PDF of $ln(I_*)$, the dashed red line indicates an independent determination of the bankfull flood. We note that I_* is only calculated for $U_* > U_{*c}$.



Figure 4.4: Map of the continental United States (Inset shows Puerto Rico) showing the locations of field sites used in this study. Field sites are colored according to logarithmic bins of drainage area. Areas possessing gravel-bedded rivers are generally well sampled, except for the Pacific Northwest where a preponderance of scientific studies are not located near USGS gages or do not report particle size in sufficient detail. The Eastern-most point within the inset is the Mameyes River.



Figure 4.5: Hydrograph processing steps using USGS stream gage 03151400 to illustrate the process. (a) IDA discharge data are used for all records prior to 2007. (b) For discharge after 2007, data are from the NWIS. (c) USGS stage discharge rating curve. (d) USGS stage and average depth channel measurements. (e) The rating curve extends beyond the depths that may be surveyed by wading. (f) Linear transformation between channel stage and average channel depth. (g) Results of the transform in (f) showing good agreement between the average depth rating curve (green line) and the average channel data (blue). Yellow diamond represents a measure of the bankfull depth taken from an independent survey showing good agreement with the transformed rating curve. Steps (a)-(g) result in a time series of average channel depth.



Figure 4.6: (a) Comparison between shifted (blue) and unshifted (red) bankfull depth and the independently surveyed bankfull depth. Shifted values have better agreement with surveyed values than the unshifted values from channel stage. (b) Final result from the hydrograph processing in Figure 5. Blue line indicates the shifted U_* data. Dashed line indicates the surveyed bankfull U_* . The unshifted data appears to be above the bankfull shear velocity over 70% of the time, which is an unphysical scenario.



Figure 4.7: Frequency magnitude distribution for $U_*/\langle U_*\rangle_{U_*>U_{*c}}$ for all stream gages. Dashed red line represents an exponential fit to normalized U_* above the threshold of motion, denoted by the thin red line. Dark blue line represents a log-bin average, light blue and gray shading represent one and two standard deviation errorbars, respectively.



Figure 4.8: Relationship between $\langle U_* \rangle_{U_* > U_{*c}}$ and U_{*bf} determined from independent surveys. The threshold of motion for the red diamonds was empirically determined from measurements of bed load transport, while the light blue circles use equation (2). (inset) The ratio $\langle U_* \rangle_{U_* > U_{*c}}/U_{*c}$ is constant for over three orders of magnitude in channel slope.



Figure 4.9: Probability density function of $ln(I_*)$ normalized by bankfull impulse. Plot represents the aggregation of normalized I_* for all rivers used in Figure 4.7. Dashed red line denotes the location of the bankfull flood.

CHAPTER 5 : Reduced channel morphological response to urbanization in a flood-dominated humid tropical environment

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Abstract:

Urbanization through the addition of impervious cover can alter catchment hydrology, often resulting in increased peak flows during floods. This phenomenon and the resulting impact on stream channel morphology is well documented in temperate climatic regions, but not well documented in the humid tropics where urbanization is rapidly occurring. This study investigates the long-term effects of urbanization on channel morphology in the humid subtropical region of Puerto Rico, an area characterized by frequent high-magnitude flows, and steep coarse-grained rivers. Grain size, low-flow channel roughness, and the hydraulic geometry of streams across a land use gradient that ranges from pristine forest to high density urbanized catchments are compared. In areas that have been urbanized for several decades changes in channel features were measurable, but were less than those reported for comparable temperate streams. Decades of development has resulted in increased fine sediment and anthropogenic debris in urbanized catchments. Materials of anthropogenic origin comprise an average of 6% of the bed material in streams with catchments with 15%or greater impervious cover. At-a-station hydraulic geometry shows that velocity makes up a larger component of discharge for rural channels, while depth contributes a larger component of discharge in urban catchments. The average bank-full cross sectional area of urbanized reaches was 1.5 times larger than comparable forested reaches, and less than the world average increase of 2.5. On average, stream width at bank-full height did not change with urbanization while the world average increase is 1.5 times. Overall, this study indicates that the morphologic changes that occur in response to urban runoff are less in channels that are already subject to frequent large magnitude storms. Furthermore, this study suggests that developing regions in the humid tropics shouldn't rely on temperate analogues to determine the magnitude of impact of urbanization on stream morphology.

5.1. Introduction

The year 2009 marks the first time in history where human populations in cities surpassed populations in rural areas (*United Nations and Social Affairs*, 2010). Consequently, a fundamental understanding of the effects of urbanization on different types of fluvial landscapes is essential for sustainable urban growth. The scientific understanding of the effects of urbanization on physical stream morphology remains limited, especially in the tropics (*Chin*, 2006). Moreover, in a comprehensive global scale review of over 100 urban stream geomorphology studies only seven were focused on the tropics, and only two were focused on changes in stream channel morphology (*Chin*, 2006). Here, we present an analysis on the effects of urbanization on stream morphology in the humid tropical region of Northeast Puerto Rico. A region that is subjected to frequent large storm events including hurricanes and tropical storms.

The effects of urbanization on stream flow and floods are generally better understood than the effects of urbanization on channel morphology. The routing of storm water across impervious surfaces and through pipes greatly reduces infiltration and groundwater recharge and consequently decreases the lag time between precipitation and stream flow response (*Leopold*, 1968, 1991; *Hollis*, 1977). It is well known that impervious cover increases the magnitude of peak discharges and decreases the duration of a flood compared to a similar non-urbanized catchment (*Dunne and Leopold*, 1978; *Leopold*, 1968, 1991; *Hollis*, 1975, 1977; *Sauer et al.*, 1983). However, the effect of urbanization on the magnitude of peak flood discharges decreases with increasing flood recurrence interval, due to the diminishing effects of infiltration and interception during large events (*Leopold*, 1968). In a review of temperate datasets on peak flood discharge and recurrence, *Hollis* (1975) determined that the largest effect of urbanization occurs for floods with recurrence intervals less than two years.

A general framework for urbanization and stream response was first put forth by *Wolman* (1967) for the region of Baltimore, Maryland. Here the initial effect of urbanization on the

stream channel network was a large pulse of fine sediment that entered the streams during the construction phase of urbanization. This was followed by a decrease in sediment supply as impervious surfaces and ground cover increased in the established urban area. Following the construction phase and the introduction of impervious surfaces the resulting increase in peak stream flow and decrease in upland sediment supply causes changes in channel morphology. Typically the sediment that was deposited in the construction phase is eroded by the higher peak discharges and the channel adjusts to a form that is typically enlarged compared to the original channel (*Wolman*, 1967).

While the Wolman model has been widely accepted (*Chin*, 2006), a difficulty with quantifying the effects of urbanization on stream morphology is that the stages of urbanization occur over variable timescales so it can take years to decades for a channel to adjust to the newly imposed flow conditions (*Wolman*, 1967; *Neller*, 1988; *Leopold et al.*, 2005; *Chin*, 2006; *Gregory*, 2006). As channels adjust to the new regime of larger magnitude flows and reduced sediment input, the most commonly reported morphological change on wadeable streams with drainage areas less than 150 km² is a change in the channel cross section (*Chin*, 2006). For alluvial streams along the Eastern Coast of the United States overall channel widening is the most reported change (*Hammer*, 1972; *Gregory et al.*, 1992; *Leopold et al.*, 2005; *Grable and Harden*, 2006; *O'Driscoll et al.*, 2009). Channel widening occurs through bank erosion, which can also introduce large quantities of sediment into the stream (*Trimble*, 1997).

Although channel widening is the most commonly reported form of channel response to urbanization, regional differences in geology and topography can lead to channel incision (*Hardison et al.*, 2009). Changes in grain size distributions on the stream bed have also been observed. In catchments located in Maryland, USA grain size within the channel became finer during the construction phase (*Wolman*, 1967). Other studies in established urban catchments report that channel bed grain size becomes coarser as lag deposits form from the selective transport of smaller grain sizes (*Konrad et al.*, 2005), or as coarse-grained materials used for bank stabilization accumulates (*O'Driscoll et al.*, 2009). In urban catchments in the Southeastern Pennsylvania region visually estimated roughness values suggest that urbanization decreased stream roughness compared to nearby rural streams (*Pizzuto et al.*, 2000). Reach-scale channel slope has been observed to decrease as a result of bed aggradation from bank erosion (*Colosimo and Wilcock*, 2007). Channel slope has also been observed to steepen as a result of channel straightening (*Grable and Harden*, 2006).

Collectively these studies indicate that the magnitude, direction, and time scales of channel change in response to urbanization vary. Moreover, responses vary with regional geology, climate, the duration and pattern of urbanization, sediment supply, and stream competence and capacity (*Wolman*, 1967; *Chin*, 2006; *Gregory*, 2006; *Hardison et al.*, 2009). Streams in different climatic regimes with higher frequencies of storm events should respond differently to urbanization. Likewise streams that are initially sediment transport or supply limited should respond differently to the effects of urbanization. Transport-limited streams that are unable to transport all the sediment introduced into the channel are likely to undergo aggradation during and after the construction phase of urbanization. Supply-limited streams that are able to transport more than the introduced sediment are likely to only experience a minor and temporary amount of aggradation during the construction phase.

This paper expands on the paucity of data on stream channel response to urbanization in humid tropical regions. Channel hydraulic and morphologic features from sites with land uses that range from established urban areas to old growth evergreen tropical forests are compared to test the hypotheses that stream channel geometry in NE Puerto Rico exhibits a smaller degree of urbanization induced alteration when compared to streams that have been studied in temperate regions. The reduced channel adjustment to urbanization is expected because these streams are sediment limited and subjected to a high frequency of sediment mobilizing storm flows. Therefore, the impact of urbanization on these streams is hypothesized to produce relatively small changes in channel morphology and at-a-station hydraulic geometry. Grain size and stream channel roughness are hypothesized to increase due to the selective transport of finer grains. The few existing studies of channel response in tropical regions suggest that channels undergoing urbanization experience an overall initial decrease in channel capacity, though to a lesser degree than that of temperate streams (*Ebisemiju*, 1989; *Odemerho*, 1992; *Gupta and Ahmad*, 1999). It should be noted that the studies by *Ebisemiju* (1989) and *Odemerho* (1992) are located in the coastal lowlands of Nigeria and do not experience large scale tropical storms such as hurricanes and typhoons, while the study of *Gupta and Ahmad* (1999) is a qualitative review of urbanization in the humid tropics.

5.2. Study site

A gradient in land use extends from the Luquillo mountains to the San Juan Metropolitan area (Figure 5.1), with urbanization increasing westward from the old growth rain forest in the El Yunque National Forest to the San Juan Metropolitan area. Precipitation averages for 32 years (1963-1995) for the region (Figure 5.1) range from 1500 mm/yr on the Western edge of the San Juan Metropolitan area to over 4000 mm/yr at the peaks of the Luquillo mountains. The region is characterized by high frequency, short duration high-magnitude precipitation events and flashy stream flow hydrographs (Schellekens et al., 2004; Smith et al., 2005). The regions flashy hydrographs are due to frequent, high-intensity rainfall, steep slopes, and efficient drainage networks (Scatena, 1989; Schellekens, 2000; Smith et al., 2005). The uplands also have a high susceptibility to landslides, and landslide derived sediment contributes a large portion of the watershed scale sediment budget (Larsen et al., 1999; Lepore et al., 2011). Estimations of sediment transport capacity, channel features, and the rapid removal of landslide dams suggest that the streams are sediment supply limited (Ahmad et al., 1993; Clark and Wilcock, 2000; Pike et al., 2010). Suspended sediment yields are high and dependent on sediment availability rather than the magnitude of peak flows (Gellis, 1993). For the water year 2003, which is typical of the longer term records, suspended sediment loads for the urbanized Rio Piedras, the mixed land use Rio Fajardo, and the forested Rio Mameyes were 3691, 1117, and 457 metric tons/km² respectively (*Diaz* et al., 2005).

The entire region of NE Puerto Rico, except the upper elevations of the Luquillo Mountains, has undergone considerable change in land use since the colonization of the island by Europeans (Scatena, 1989). By the 1950s up to 94% of the island, and most of the lowland areas studied here were deforested (Birdsey and Weaver, 1987). Since the 1950s the entire island has seen considerable reforestation and expansion of the San Juan metropolitan area (Helmer, 2004). One effect of these land use changes is an increase in stream sediment storage and shallowing of some alluvial reaches at the base of the Luquillo Mountains (Clark and Wilcock, 2000). Analysis of stream flow quantity and frequencies over the past 50 years of reforestation found there was no simple or significant relationship between reforestation and stream flow in 12 meso-scale (23–346 km²) catchments (Beck et al., 2013). Both Beck et al. (2013) and Wu et al. (2007) showed similar local decreases in stream flow with reforestation on the Eastern coast of the island, however both studies avoid heavily urbanized areas. Today, urban and suburban development predominantly occurs in the coastal plains at the base of the mountain ranges. Urbanization has had an overall negative impact on water quality in the San Juan Metropolitan area and results in higher concentrations of Phosphorus, Sodium, lower dissolved Oxygen, and high quantities of fecal coliform (Santos-Roman et al., 2003; Diaz et al., 2005). However extensive sampling of aquatic organisms indicates that urbanization has not had a large impact on the diversity or abundance of fish populations in the San Juan Metropolitan area (*Ramirez et al.*, 2009).

The majority of the urban field sites for this study reside in the Rio Piedras watershed (Figure 5.2(f)), which is one of several rivers draining the greater San Juan Metropolitan area. A majority of the Rio Piedras drainage network has been directly altered by human interaction through channelization, or by being routed through canals (*Lugo et al.*, 2011). There are also significant numbers of storm drains that route directly into the Rio Piedras River. In this paper we explore the effects of urbanization on the physical morphology of self-forming stream channels and avoid field sites that have recently been altered by direct
human actions.

5.3. Methods

This study evaluated the influence of urbanization by comparing the hydrology and channel morphology of streams that drain catchments with a range of impervious area. Regional hydrology and peak flows were analyzed to determine the impact that urbanization has had on the magnitude and frequency of stream flow. Grain size distributions were compared to test if larger flood peaks and an aging urban establishment result in coarser bed surface sediment. Stream bed roughness was computed to assess if urbanization increased channel roughness. At-a-station hydraulic geometry for U.S. Geological Survey (USGS) gaging stations was computed to further quantify morphological and hydraulic changes associated with urbanization in the region. Longitudinal hydraulic geometry for reference levels relevant to the initiation of sediment transport and flow frequency were also assessed to quantify the influence of impervious cover on channel morphology.

Data from survey reaches presented in this paper consist of 10 reaches at USGS stream gaging stations and 40 reaches located in the surrounding region (Figure 5.1). All of the reaches were in watersheds that primarily drain volcaniclastic rocks with deeply weathered clay soils. The reaches were surveyed during the summers of 2009-2011 and located using a Trimble GeoXH differential GPS. Survey reaches were chosen to capture the effects of land use change across a large range of drainage areas (0.028 - 40.33 km²) within the region (Figure 5.2).

Surveyed stream reaches were straight uniform alluvial reaches where the channel could self adjust its geometry. Each surveyed reach was 6-10 channel widths long. Areas where bedrock expressed significant control on the reach geometry or reaches that showed signs of recent direct anthropogenic alteration were not selected. Several channelized reaches were surveyed to provide a comparison to the unchannelized urban and rural reaches. All surveyed reaches were situated away from bridges due to the strong control on channel geometry and erosion that bridges can introduce (Douglas, 1985).

Longitudinal hydraulic geometry

At each study reach the cross section, grain size, and longitudinal profile were measured. Longer reaches were surveyed using a Sokkia Total Station Laser Theodolite (set 530) for increased accuracy, while smaller reaches (less than 60 meters in length) were surveyed using an Impulse 200 Laser Rangefinder with Mapstar Compass Module II attachment. In order to minimize intra site variability in channel geometry measurements, three cross sections at each reach were surveyed and their metrics averaged. These cross sections were located at the upper, middle, and lower sections of the study reach and extended from vegetated hill slope to vegetated hill slope or from floodplain surface to floodplain surface when a floodplain was evident. For each cross section survey, measurements were taken at a maximum spacing of one meter in the cross stream direction. Longitudinal profiles were surveyed with at least ten points in the channel thalweg over a distance of approximately ten times the channels width. Width (W), depth (h), and cross sectional area (A_{xs}) were calculated from each cross section. The location of morphologic breaks (Figure 5.3) and riparian features (*Pike et al.*, 2010) were also recorded. Values reported in this paper for W, h, and A_{xs} represent the average of the three cross sections for a given reach at the specified reference depth. A linear regression best fit line was fit to the longitudinal profile to determine the slope of the reach. Grain size distributions were determined using a random walk pebble count (Wolman, 1954), where all grains smaller than 2 mm where recorded as 2 mm. Anthropogenic debris of a size and density large enough to behave as bed load, such as concrete and asphalt, were measured as part of each pebble count.

The surface roughness of the channel at low flow was calculated for each reach from the measured reach slope and a measure of average low-flow reach velocity. The velocity measurement was made using the salt dilution method at each non gaged site. The salt dilution method was chosen over other methods of measuring velocity due to the shallow flow depths,

and coarse grain size in many sites. Flow velocity for the USGS gauged reaches was calculated from USGS discharge records for the same time period of the field survey.

Stream channel skin roughness was computed using the Manning-Strickler approach. Roughness was measured at shallow flow conditions to minimize the impact of frictional resistance of the banks. The Manning-Strickler approach is based on the premise that frictional resistance for a steady uniform flow should be equal and opposite to the downstream component weight of the water column. The Manning-Strickler relationship can be reformulated directly to solve for the roughness coefficient Mannings n:

$$n = \frac{h^{2/3} S^{1/2}}{U} \tag{5.1}$$

where h is the flow depth, S is the channel slope, and U is the flow velocity. Here we use a reach slope and reach average velocity as described in the methods. Darcy-Weisbach, Chezy, and Keulgan friction factors were also calculated and compared. They all provide similar results as the Manning-Strickler relationship. For a detailed discussion of roughness relationships see *Ferguson* (2007).

Longitudinal changes in hydraulic geometry and channel morphology were compared at two reference depths. Channel geometry is commonly compared at the bank full flow depth (*Hammer*, 1972; *Galster et al.*, 2008; *Faustini et al.*, 2009). However a topographic expression of bankfull flow is often obscured or absent from steep gradient channels in this region (*Pike and Scatena*, 2010), and steep gradient mountain streams elsewhere (*Wohl et al.*, 2004). Furthermore significant changes or differences may occur at flows other than bankfull. Therefore we compared geomorphic and hydraulic features at two different reference levels (Figure 5.3(a)): the height on the bank at which trees are established, and the water depth required to initiate motion of the median grain size (D_{50}). The height on the channel bank at which trees are established (h_{Tree}) has been shown in this region to be inundated at a frequency that is comparable to bankfull flow in adjacent reaches with defined floodplains (*Pike and Scatena*, 2010). This flow also has a median recurrence interval of approximately 55 days and corresponds to the effective discharge of the channels when magnitude and frequency of flow and sediment transport are considered.

The water depth required to initiate motion of the D_{50} was used as a second reference level. A reference level based on sediment transport is relevant because morphological change within channels occurs through the erosion and deposition of sediment. Changes within a streams catchment that affect the hydrology and transportability of sediment should be reflected in the geometry of the channel at flow depths relevant to mobilizing sediment. The critical depth for the D_{50} is supported as gravel streams are well known to adjust their width and depth such that the active channel is slightly above the critical depth to initiate movement of the D_{50} (*Parker*, 1978; *Dade and Friend*, 1998; *Parker et al.*, 2007). The concept of equal mobility in gravel streams (*Wiberg and Smith*, 1987; *Parker and Toro-Escobar*, 2002) also guided the choice of using the critical depth for the D_{50} as a reference level. Initiation of motion for sediment occurs as the shear stress on the bed reaches a critical value and can be estimated using the Shields equation:

$$\tau_* = \frac{\tau_b}{(\rho_s - \rho)gD} \tag{5.2}$$

where τ_* is the dimensionless shear stress or Shields stress, τ_b is the basal shear stress, ρ_s is the density of sediment (2650 kg/m³), ρ is the density of water (1000 kg/m³), g the acceleration due to gravity (m/s²), and D is the median axis of a representative grain size (m). The basal shear stress under steady and uniform flow can be approximated as:

$$\tau_b = \rho g h S \tag{5.3}$$

Equation 5.3 can be substituted into Equation 5.2 and rearranged to give:

$$h_{cr} = \frac{\tau_{*cr}(\rho_s - \rho)D}{\rho S} \tag{5.4}$$

where h_{cr} is the critical depth (m), and τ_{*cr} is the critical Shields stress for initiation of

motion of sediment with grain size D. The critical Shields stress typically ranges from 0.03 to 0.073, and for this study is taken to be 0.05 which is the average of 128 studies (*Buffington and Montgomery*, 1997). An empirical relationship developed by *Mueller et al.* (2005) shows that the critical Shields stress changes with increasing grain size and slope. When applied to these reaches the corrected critical Shields stress varied only slightly from the constant value of 0.05, and therefore was not used in this study.

Equation 5.4 was solved for h_{cr} for the D_{50} , which we refer to as the critical D_{50} reference depth (h_{crD50}) . We recognize that not all field sites in this study will begin transporting sediment at a Shields value of 0.05, however since Shield stress is not constant and is dependent on the state of the bed, with the possibility of changing from flood to flood (*Frey* and *Church*, 2011) the use of 0.05 represents a middle value.

The height at which morphologic breaks (Figure 5.3(a)) in each channel cross section occur were compared to the reference flow depths. Morphologic breaks are often coincident with a form of vegetation or level of soil development, which has been shown to be at a consistent flow frequency in these types of channels (*Pike and Scatena*, 2010). Additional support for the use of the critical D_{50} reference depth is that 83% (35 of 42) (Figure 5.3(b)) of the field sites possess a morphologic break within 10 cm vertically of the predicted h_{crD50} .

Expected change ratios were calculated following the method as first applied by Hammer (1972) for urban streams in Southeast Pennsylvania. In this method a relationship between a channel geometry metric and drainage area (DA) is determined for minimally impacted (I_{low}) sites through best fit regressions. The determined relation can then be used to predict the expected channel geometry in the absence of impervious cover. The resulting expected change ratio is the observed channel geometry over the predicted channel geometry. In the present study we used the average yearly rainfall for each watershed multiplied by each sites drainage area as the independent predictor. This accounts for changes in annual precipitation and drainage area between sites.

At-a-station hydraulic geometry

The at-a-station hydraulic geometry relationship was calculated for each USGS stream gaged reach using the measured values of W, A_{xs} , U, and discharge (Q) reported by the USGS. Values for the average cross section depth (h) were determined by dividing cross sectional area by width. Leopold and Maddock (1953) found that active channel discharge is related to W, h, and U by: $W = aQ^b$, $h = cQ^f$, $U = kQ^m$ respectively. Therefore the sum of the exponents (b, f, m) and the products of the coefficients (a, c, k) should both equal one. Here, as in similar studies, the exponents and coefficients were determined by a least squares log-log regression of USGS data (available online at waterwatch.usgs.gov). All values annotated with poor in the published dataset were discarded and only datasets where the sum of the exponents and product of the coefficients approached one were used. For several hydraulic geometry records there are two clearly separate relations represented in the published data, which indicates differences in the location of where the measurements were made. In these cases we fit least squares logarithmic regression lines to each grouping of the data and used the dataset whose exponents most closely sum to one and possess the highest R^2 value.

Stream flow

Stream flow duration curves of 15 minute instantaneous discharge for a ten year interval were computed to assess changes in flow frequency across the region (Figure 5.4). All flow data are from USGS stream gages and covered the same time interval, 2000 through 2009, except for two stations which only had eight years of data (2002 through 2009). Due to the short duration of most flood events the 15 minute instantaneous discharge was used instead of daily average discharge. The flow duration curves were also standardized by drainage area to account for differences in basin size.

Land use, and geographic analysis

Drainage area, road density, land use, geology, and precipitation were calculated for each field site from digital maps of the region using ARCgis (version 9.3). Drainage area was calculated for each catchment as all the area upstream from the field site location using a flow routing algorithm on one arc-second (approximately 30 meters) digital elevation maps from the 2010 National Elevation Dataset (*Gesch et al.*, 2002). Land use is derived from a dataset produced by the USGS Land Cover Institute and is divided into seven categories (*Homer et al.*, 2004): Developed Open (D_o), Developed Low (D_L), Developed Medium (D_m), Developed High (D_h), Evergreen, Pasture, and Grassland.

The total impervious area in each catchment is a combination of the four developed land use categories. Each developed land use category as defined in the dataset by (*Homer* et al., 2004) represent a range of impervious surfaces types and spatial extent, such that the total impervious area for each land use type is a fraction of the total area it occupies. Impervious area of each catchment was calculated from a linear combination (Equation 5.5) of the developed land use types and a coefficient that specifies the fraction of developed land use type that is impervious. We combined road length (R_L) and impervious area by assuming an average width of 10 meters per road.

Impervious Area =
$$0.1D_o + 0.35D_L + 0.65D_m + 0.9D_h + 0.01R_L$$
 (5.5)

Coefficients in equation 5.5 represent the median value of the range of impervious cover for each class of impervious cover as defined by the NLCD Land Cover Class Descriptions (*Homer et al.*, 2004). For example, land cover in the D_L category ranges from 20-49% impervious cover (*Homer et al.*, 2004). Use of the median value of 0.35 for D_L is unlikely to over or under predict the impervious cover in a given grid cell. Roads represent a small fraction of total catchment area, but remain non trivial sources of anthropogenic stream alteration (*Sampson*, 2000; *Nelson and Booth*, 2002). We acknowledge that 10 meters is a conservative estimate for road width and we include road area in the sum of impervious area for completeness. Roads represent a large percent of the impervious area calculated for rural sites, and a small percent of impervious area for urbanized sites. Roads represent a small fraction of total drainage area for all sites.

Field sites were grouped into three categories of impervious cover: low (I_{low}) , medium (I_{med}) , and high (I_{high}) . The three categories correspond to impervious cover ranges of <10%, 10-20%, and >20% respectively and are similar to groups defined by *Schueler et al.* (2009). The sample size for I_{low} , I_{med} , and I_{high} are 21, 9, and 13 sites respectively, and fall within the range of sample sizes of comparable studies (examples: *Hardison et al.*, 2009; *Pizzuto et al.*, 2000; *Shepherd et al.*, 2011). Relationships were also compared using percent forest cover in the watershed instead of percent impervious cover. The results were similar when using impervious cover or forest cover, and percent forest cover is well anti-correlated to percent impervious area (Pearson correlation coefficient of -0.76). We report results using impervious cover to build on the existing body of impervious cover and urbanization literature.

Total average yearly precipitation data was extracted from the 30 year normal PRISM dataset created by *Daly et al.* (2003). Total annual precipitation of each catchment is the sum of all grid cells of the 30 year average precipitation dataset within a field sites catchment. Average yearly precipitation represents the total precipitation divided by the total number of gridded cells (grid size is 0.2025 km^2) within the field sites catchment. Due to the presence of a precipitation gradient in the region (Figure 5.1) we calculated the total average yearly precipitation for each field site by multiplying drainage area (m²) and the average precipitation (m/yr) for the upstream area. This metric is the 30 year average total annual water input into the drainage area, which we refer to as the total precipitation flux (T_{PF}).

Statistical tests used in the following analysis are the Mann-Whitney-Wilcoxon and Kruskal-Wallis tests. Statistical tests were always performed on data that has been standardized by either dividing by DA, or T_{PF} . Results for the Mann-Whitney-Wilcoxon two-sample ranked-sums test are reported as (W_{stat}, P_{val}) where W_{stat} is the Mann-Whitney U statistic; we report the smaller of the two rank sums, and P_{val} is the associated P-value. Results for the Kruskal-Wallis test are reported as $(H_{stat}, .d.f., P_{val})$ where H_{stat} is the Chi-squared statistic, and d.f. the degrees of freedom of the Chi-squared distribution. Sample sizes where relevant are reported in the associated figures. In the following analysis statistical test results with a P-value of 0.05 or less, 95% confidence level, are considered significant. All statistical analysis were performed using R version 2.10.1.

5.4. Results

Regional trends

Field sites cover a range of drainage areas (.07-39.49 km²), slopes (0.001-0.049), and precipitation (1584-3721 mm/yr) (see Table 5.1 for averages and range for each impervious group). Average annual precipitation decreases outward from the Luquillo Mountains and increases with elevation. Reach slope decreases as drainage area increases across all sites (R²=0.59, P_{val} =<0.001), but there is considerable scatter in the relationship. Grain size also decreases with increasing drainage area, although several streams maintain a fairly coarse grain size well into the coastal plain, and in some cases cobbles are observed at the mouth of the river. For all sites W, h, and A_{xs} increase with increasing drainage area.

Hydrology

The unit flow magnitude observed at high and low flow frequencies are generally larger for I_{low} streams compared to I_{high} streams (Figure 5.4). However, unit flows at the 1% frequency are comparable across all land uses and sites. Comparison of the annual peak flow time series data for the last 30 years reveals that I_{low} streams have higher yearly peak flow than the I_{high} streams when accounting for differences in catchment size. These differences are partly caused by the precipitation gradient across the region (Figure 5.1) and indicate that urbanization in this region does not necessarily result in increased peak flows compared to adjacent rural areas. They also indicate that differences in precipitation need to be accounted for when comparing channel responses between sites (see below).

Grain size

Grain size distributions for all sites are poorly sorted with considerable overlap between sites grain size distributions. Sorting also remains fairly constant with increasing drainage areas. Grain size metrics were divided by DA to account for varying catchment sizes. The Kruskal-Wallis test of statistical significance between the variability of grain sizes for all three impervious classes indicates there was no statistical difference between the medians in D_{16} (H_{stat} =1.24, 2 d.f., P_{val} =0.53), D_{50} (H_{stat} =0.63, 2 .d.f., P_{val} =0.72), or D_{84} (H_{stat} =0.93, 2 d.f., P_{val} =0.62). We further tested for significance whether the grain size populations differ between the Ilow and Ihigh groups using the Mann-Whitney-Wilcoxon test for D_{16} (W_{stat} =158, P_{val} =0.62), D_{50} (W_{stat} =176, P_{val} =0.54), and D_{84} (W_{stat} =184, P_{val} =0.38). Sample sizes for each group are labeled in Figure 5.5. Statistical results are similar when comparisons are standardized by T_{PF} .

Comparison of the grain size distributions does reveal that sites with medium and high impervious area have a larger abundance of sand sized particles than sites with low impervious area (Figure 5.5(Inset)). A Mann-Whitney-Wilcoxon test indicates that the difference in percent sand between the high and low impervious classes are statistically significant at the 95% level (W_{stat} =63, P_{val} =0.02). Multivariate linear regression models with the combination of T_{PF} , slope, and impervious cover explained 61% of the variance in the fraction of sand. However, the amount of impervious cover in the watershed only accounts for 10% of the explained variance.

All but the most remote streams in the region have some form of anthropogenic debris present within the active channel (Figure 5.5(Inset 2)). The anthropogenic debris recorded in pebble counts are of a size and density that is likely to behave as bed load. Streams with greater than 10% impervious cover have at least 1% of the bed cover composed of concrete, automobile parts, and other non-easily erodible material of anthropogenic origin. For the collected data the percent of anthropogenic clasts increases linearly with impervious cover. Streams with greater than 15% impervious cover have an average of 6% of the bed material of anthropogenic origin. Streams with 40% or more impervious cover have an average of 11% and a maximum of 14% of the bed material of anthropogenic origin. It should be noted that grain size distributions in the urbanized streams can be very dynamic and can change from storm to storm depending on the input of fine sediment. For example, a site on the Rio Piedras had a dramatic shift in grain size between the summers of 2009 and 2010 where the D_{50} changed from 79 mm to 30 mm respectively. In contrast, comparisons of grain sizes between the forested sites measured in this study and in *Pike* (2008) indicate grain size distributions are similar between the studies and that grain size is relatively constant in the forested reaches.

Stream channel roughness

The roughness metric Mannings n was standardized by the sites respective T_{PF} value to account for differences in both DA and precipitation for all statistical tests. Roughness values from Mannings n show a decrease in roughness with increasing T_{PF} (Figure 5.6). Mann-Whitney-Wilcoxon tests indicate there were no significant differences between the I_{low} and I_{high} impervious groups for the median of Mannings n ($W_{stat}=124$, $P_{val}=0.53$). The variance for Mannings n is not statistically significant at the 95% confidence level (Kruskal-Wallis test: Mannings n $H_{stat}=1.18$, 2 d.f., $P_{val}=0.55$).

At-a-station hydraulic geometry

At-a-station hydraulic geometry exponents (see Table 5.2) for the relationships between depth (f), width (b), and velocity (m) determine the degree to which each component changes with increasing discharge. The results indicate that changes in width comprise a small component of discharge for all streams and never contributes more than a third of

the discharge (Table 5.2). Velocity has the largest rate of change with increasing discharge for streams with low amounts of impervious cover. Depth has a comparatively larger rate of change with increasing discharge for streams with greater amounts of impervious cover. A gradient in impervious cover is apparent in Figure 5.7 such that the hydraulic geometry exponents have a larger velocity component for low impervious cover streams to an increasing depth exponent for more urbanized catchments. Boxed urbanized channels represent channelized reaches, which plot between the rural and highly urbanized catchments. Comparing this hydraulic geometry data with measurements for mixed land use basins located in Southern Puerto Rico that are predominantly underlain by limestone and volcaniclastic lithologies (*Lewis*, 1969) supports these observations and indicates that the trends are not likely climatic or geologic.

Longitudinal hydraulic geometry

Longitudinal hydraulic geometry relationships are used to analyze the shape of the channel at a specified reference height as a function of the field site's location in the catchment and total impervious area. The analysis indicates that the first order predictor of W, h, and A_{xs} for any stream in the region is T_{PF} . The combination of DA and P as separate predictors explains a comparable amount of variance in regression analysis as T_{PF} . Multivariate regressions of log transformed variables also indicates that T_{PF} provides the best fitting model and provides the relation explaining the highest amount of variance. Multiple linear regression results (Table 5.3) show that with two exceptions, between 70 and 81 percent of the explained variance in W, h, A_{xs} at both reference levels can be explained by T_{PF} alone. For h at both reference levels no more than 54% of the variance can be explained with multiple independent predictor variables.

Expected change ratios for W, Axs, and W/h at both reference depths are presented in Figure 5.8(a), 5.8(b), and 5.8(c) respectively. Coefficients of determination for the best fitting functions for W and A_{xs} that were used to calculate the change ratios range from 0.85 to 0.92 and average 0.9 (Table 5.4) and indicate that the ratios are based on strong relationships. Coefficients of determination for W/h ratios are lower (Table 5.4) than those for W and A_{xs} , because h exhibits more variability. Change ratios for W/h were calculated from two separate predictive relationships for both W and h for the h_{crD50} reference level due to a poor relation between T_{PF} and W/h. For the h_{Tree} reference level W/h change ratios were calculated using a single relationship between W/h and T_{PF} . The channel change ratio for W shows the largest range of values at the h_{crD50} reference level (Figure 5.8(a)). Kruskal-Wallis tests for each reference level indicate that none of the impervious classes are significantly different at the 95% confidence level (Kruskal-Wallis test results: h_{crD50} $H_{stat}=2.14, 2 d.f., P_{val}=0.34; h_{Tree} H_{stat}=1.04, 2 d.f., P_{val}=0.59$). None of the observed differences in W for the three categories of impervious area within each reference level are statistically significant at the 95% confidence level when using a Mann-Whitney-Wilcoxon test.

The channel change ratio for A_{xs} has the largest range of values at the h_{crD50} reference level (Figure 5.8(b)). Kruskal-Wallis tests for each reference level determine that the group medians are not statistically significantly different at the 95% confidence level (Kruskal-Wallis test results: $h_{crD50} H_{stat}$ =3.13, 2 d.f., P_{val} =0.21; $h_{Tree} H_{stat}$ =2.17, 2 d.f., P_{val} =0.34). A Mann-Whitney-Wilcoxon test for all impervious groups within each reference level yields results similar to the statistical analysis performed on the W change ratios; none of the observed relationships are statistically significant at the 95% confidence level. For comparison surveyed channelized reaches have an average change ratio of 3.3 at the h_{Tree} reference level.

The channel change ratio for the width depth ratio (W/h) suggests that at the h_{crD50} reference level (Figure 5.8(c)) the I_{high} category has a larger range of W/h and higher median W/D ratio. Interestingly the W/h ratio for the h_{Tree} reference level has the smallest range of values and the lowest median W/h ratio. These results are similar in trend to the W change ratio in Figure 5.8(a). Kruskal-Wallis tests for each reference level determine that the

group medians for the h_{Tree} but not the h_{crD50} reference level are statistically significantly different at the 95% confidence level (Kruskal-Wallis test results: $h_{crD50} H_{stat}=1.72, 2 d. f.$, $P_{val}=0.42$; $h_{Tree} H_{stat}=7.03, 2 d. f.$, $P_{val}=0.03$). A Mann-Whitney-Wilcoxon test for all impervious groups within each reference level yields results similar to the statistical analysis performed on the W and A_{xs} change ratios; none of the observed individual relationships are statistically significant at the 95% confidence level.

5.5. Discussion

First order trends in channel cross section morphology in this area are well described by precipitation and drainage area and have inter site variations similar to other regions (*Hack*, 1957; Wohl and Merritt, 2008; Pike et al., 2010). Analysis of flow duration curves and peak flow magnitude indicate that in this environment of large and frequent rains, changes in impervious cover result in only minor changes in peak stream flow. The small changes in stream flow are likely due to the rapid generation of storm flow throughout the region, which is a result of steep slopes and efficient drainage networks (Schellekens et al., 2004). Moreover when streams of different sizes and locations are considered, differences in surface hydrology between impervious groups are not substantial and can be largely explained by drainage area and precipitation differences. Beck et al. (2013) reached similar conclusions when looking at the influence of 50 years of reforestation on Puerto Rican streams. Apparently the combination of frequent sediment mobilizing events, watershed, and climatic heterogeneity reduce and or obscure hydrologic and morphologic responses to land cover change in this environment. Nevertheless, the above results do indicate that when differences in annual precipitation and drainage basin area are considered, urbanization has influenced the channel morphology and grain size distributions of these streams. Comparison of historical photographs of the lower Rio Piedras, taken in 1936 and 2002, also show drastic reductions in channel sinuosity via channelization (Lugo et al., 2011).

The addition of anthropogenic debris is a direct impact of urbanization on grain size distributions in the region, with on average 6% of the bed material derived from anthropogenic origin for sites with greater than 15% impervious cover. The measured increase in fine sediment is likely a result of ongoing construction in the urban areas. This is further supported by satellite images throughout the last decade of the region showing that many of the I_{high} sites catchments are still actively being developed. While the differences in fine sediment observed between I_{high} and I_{low} sites is statistically significant, the difference between their medians only represents a 10% increase in sand, which is unlikely to cause large changes in channel depth or morphology. Increased channel sand has been observed to cause temporary channel aggradation, however widespread channel aggradation has not been observed in these channels. Increased sand can also increase the transportability of coarser sediment (Wiberg and Smith, 1987; Wilcock, 1998; Wilcock et al., 2001). Modeling results from Gasparini et al. (1999) suggest that sand enhanced transport of larger grains occurs when the bed is greater than 40% sand. Thus the 10% increase in sand (Figure 5.5(inset)) observed here is not enough to significantly alter the transport of coarse-grained sediment or change channel hydraulics or morphology. Furthermore including the amount of sand in the channel does not contribute any explanatory power to multiple linear regression models of channel geometry with other independent factors (Table 5.3). Therefore, while urbanization did result in measurable differences in the grain size distributions, they are not large enough to result in measurable differences in the transport capacity of the channel or result in channel aggradation on the scale that has been observed in some temperate streams.

Measurements of Mannings n further indicate that urbanization does not result in measurable differences in channel roughness in these streams (Figure 5.6). Differences in roughness between impervious cover groups are not statistically significant. High sediment yields (*Gellis*, 1993) and observed changes in grain sizes indicate these landscapes are still actively supplying sediment to the urban streams, but that this supply has not altered channel roughness or exceeded the streams transport capacity. The steep gradient and coarse-grained nature of the rivers in Northeast Puerto Rico seems to act as a buffer to changes in the overall bed skin roughness.

At-a-station hydraulic geometry reveals that velocity and to a lesser extent depth make up the largest components of discharge in these streams (Figure 5.7). The depth exponent also accounts for a larger part of stream discharge in urbanized streams. This is supported by field observations and W/h ratios of I_{high} sites having slightly narrower, deeper, and more box-like channels. However, of the three channel geometry metrics that were compared, depth commonly had the poorest model fits here and in similar studies (*Cianfrani et al.*, 2006). This is apparently because of high internal variation in channel depth and because channel depth and h_{crD50} represent dynamic channel processes that can change with the frequent sediment mobilizing storms in the region. The high variability in channel depth required more field sites to be excluded from the predictive W/h change ratio equations (Table 5.4); however the relationships yield interesting insight into the channel shape. The W/h change ratio (Figure 5.8(c)) shows that the I_{high} channels are wider at the lower reference level (h_{crD50}) and become narrower than the I_{low} channels at the higher reference level (h_{tree}) . This trend of a wider channel at the base and narrower at higher flows supports the larger contribution to discharge from depth seen in the At-a-station hydraulic geometry (Figure 5.7). This wider low-flow height and minimal widening of the channel as flow increases when compared against the I_{low} sites suggests that a more rectangular shape supporting the observation of the I_{high} sites being more "box-like". The minimal explanatory power provided by impervious cover in predicting W, and A_{xs} (Table 5.3) further suggest that they are primarily controlled by climate and drainage area and that h is not a reliable indicator of channel change for this region. The change ratios for W(Figure 5.8(a)) and A_{xs} (Figure 5.8(b)) also suggest that increased impervious cover in the watershed does not dramatically change the widths and cross-sectional areas of these channels. Moreover, all the change ratios and the combined results of this study suggest that at the time of measurement impervious area and urbanization has not drastically changed the geometry of self forming streams in NE Puerto Rico.

The timescales over which streams respond to urbanization post construction are highly variable ranging from years to multiple decades (*Wolman*, 1967; *Leopold et al.*, 2005). Con-

sidering the conceptual framework of Wolman (1967) where urbanized streams are hypothesized to reach a quasi-stable channel form determined by the post development flow regime. The question arises though whether the results presented here represent adjusted conditions expected to persist or are the results indicative of early partial adjustment and continued channel enlargement. A majority of the I_{high} streams are located in the Rio Piedras catchment which experienced a large boom in population growth and construction during the 1950s and 60s and slower rates of development since then (Lugo et al., 2011). From satellite photographs it is clear that some construction is ongoing, but the concentrated boom in development suggests that the Rio Piedras has been in the post construction phase of Wolman's conceptual framework for 30 to 40 years. The I_{med} sites are located in catchments with beginning and ongoing development. The trend of increasing A_{xs} (Figure 5.8(b)) and decreasing W/h ratio (Figure 5.8(c)) across land use at the H_{tree} reference level suggests a sequence of adjustment concurrent with the amount of development. Given the time since the major development in the Rio Piedras we suspect that either the channels have reached a new quasi-stable form or the rate of change due to urbanization has drastically slowed. Given the frequency of high-magnitude flows in the region, the time since major development in the Rio Piedras, and the overall small changes to channel morphology it is likely that the I_{high} channels are adjusted to the current conditions.

In summary the streams in the Northeast region of Puerto Rico have experienced a smaller degree of urban induced channel alteration when compared to data for the rest of the world reported in *Chin* (2006). The following comparisons are for metrics at the h_{Tree} reference level and bankfull metrics for the world dataset. Change ratios of stream width for the I_{high} Puerto Rican sites averaged 1.0 and ranged from 0.62 to 1.56. The channelized reaches in Puerto Rico were on average 1.43 times wider than comparable I_{low} channels. In contrast, the world average increase in the width of self-forming channels is 1.5 and can be as large as 7.4. The average change ratio for cross sectional area for unchannelized urban sites in this study is 1.5, while the world average is 2.5 (*Chin*, 2006). The channelized Puerto Rican reaches were on average 3.3 times larger than comparable I_{low} streams when accounting for

drainage area and precipitation.

5.6. Conclusion

The effects of urbanization on streams for this region are measurable, but muted compared to the results seen in most alluvial streams in temperate climates. As our hypothesis predicted, for this region increases in impervious cover does not result in large or statistically significantly different stream channel widths and cross sectional areas when differences in average annual precipitation and drainage area are accounted for. Channels that drain watersheds with high amounts of impervious cover have similar widths, but are slightly larger than comparable streams draining comparable rural or forested areas. Channel beds are dynamic, but there were no measureable differences in channel roughness that is directly relatable to impervious cover. At-a-station hydraulic geometry does indicate that depth represents a larger component of discharge in I_{high} streams. Both at-a-station hydraulic geometry and longitudinal hydraulic geometry suggest that urban channels are more boxlike and slightly larger than comparable channels in forested areas. The observed muted response is apparently the result of frequent high-magnitude storms that periodically flush fine sediment introduced by construction from the already supply-limited system, resulting in limited aggradation. The muted response is also due to a smaller hydrological change from the addition of impervious cover. This is due to steep slopes with saturated soils that have high infiltration rates (Schellekens et al., 2004), and a strong precipitation elevation gradient.

Overall this study indicates that urbanization has had minor impacts on the morphology of streams that have not been directly altered through channelization. This minimal alteration of channel morphology with urbanization provides a partial explanation to the observation that urbanization has not drastically altered the assemblage of aquatic species in these streams (*Ramirez et al.*, 2009). The results also indicate that the hydrological and geomorphic effects of impervious cover on channel morphology are not drastic in sediment limited streams that are subject to frequent high-magnitude flows. These results further indicate that the morphology of coarse-grained streams subject to frequent sediment transporting flows is both resistant and resilient to the addition of impervious surfaces. Furthermore, this study shows that under these conditions the conceptual model by *Wolman* (1967) is essentially correct, but that the magnitude of urbanization induced alteration is highly dependent on regional climate, and whether the river is sediment transport-limited or supply-limited. Finally the study indicates that humid tropical regions with similar climatic and elevation gradients as North East Puerto Rico shouldnt rely on temperate analogues to determine the magnitude of impact of urbanization on stream morphology.

Table 5.1: Average and range of values of selected variables for all field sites by impervious cover groups. Values in parenthesis represent the range of the data. Bold values represent the median. The median was chosen over the average when it better represented the central tendency of the data. Impervious and Evergreen Forest cover represent the percent of drainage area designated as that land use.

Site groups	Slope	Drainage	Precipitation	D_{50}	Impervious	Evergreen
(Impervious $\%$)(n)		area (km^2)	(mm/yr)	(mm)	(%)	(%)
I_{low} (<10%)	0.007	11.21	2819	35	5	68
(23)	(0.001-0.02)	(0.22-38.9)	(1940-3721)	(3-250)	(2-9)	(15-99)
$I_{med} (10-20\%)$	0.008	6.45	2152	20	13	46
(9)	(0.003-0.05)	(0.07-32.4)	(1587-2540)	(10-110)	(10-19)	(22-62)
$I_{high} (> 20\%)$	0.009	10.10	1885	25	33	25
(13)	(0.001-0.021)	(0.81-39.5)	(1584-2174)	(5-92)	(22-55)	(1-49)

Table 5.2: Hydraulic geometry coefficients and exponents for USGS stream gage data in the Northeast region of Puerto Rico. Exponents f, b, m, and coefficients c, a, k represent variables h, W, and U respectively. Drainage area and impervious cover are included for comparison, such that sites with the highest impervious cover are at the bottom of the table.

USGS name	gage	f	b	m	c	a	k	b+f+m	ack	DA	Impervious
	#									(km^2)	area $(\%)$
Mameyes nr	50065500	0.34	0.3	0.36	0.33	10.0	0.3	1.00	1.00	17.7	1.6
Sabana											
Blanco	50076000	0.18	0.13	0.69	0.24	16.73	0.25	1.00	0.99	31.40	1.8
Sabana	50067000	0.29	0.18	0.53	0.35	12.88	0.23	1.00	1.00	12.48	1.9
Fajardo at	50070900	0.18	0.23	0.59	0.28	13.1	0.28	1.00	1.01	23.93	1.9
Paraiso											
Fajardo at	50071000	0.21	0.32	0.47	0.25	14.1	0.28	1.00	0.99	38.85	2.6
Fajardo											
Grande	50064200	0.29	0.18	0.53	0.34	10.93	0.27	1.00	1.00	19.09	2.7
Canovanas	50061800	0.25	0.16	0.59	0.32	11.44	0.27	1.00	1.00	25.89	4.4
Mameyes	50066000	0.26	0.18	0.55	0.32	14.97	0.21	0.99	1.01	33.70	6.4
Espiritu Santo	50063800	0.4	0.07	0.51	0.32	19.67	0.16	1.00	0.99	22.30	7.3
Dos Bocas	50059210	0.35	0.25	0.41	0.31	10.2	0.32	1.00	1.01	32.36	9.7
Canas	50058350	0.36	0.18	0.46	0.26	7.0	0.55	1.00	1.00	19.49	13.5
Piedras nr	50048770	0.44	0.29	0.27	0.27	10.4	0.35	1.00	1.00	19.37	23.8
El Senorial											
Caguitas	50055225	0.33	0.32	0.35	0.28	10.82	0.31	1.00	0.94	31.10	26.6
Piedras nr	50049100	0.45	0.09	0.46	0.30	11.62	0.29	1.00	1.00	39.49	41.1
Hato Rey											
Josefina	50049310	0.53	0.03	0.44	0.12	14.53	0.57	1.00	1.00	49.18	73.1

Table 5.3: Reference levels (Reference), dependent variables, best fitting multiple regression models. V_{Imp} is the variance explained by impervious area is the amount of variance explained by the addition of impervious cover to the model as an independent predictor.

Reference	Variable	Model	\mathbf{R}^2	F	Р	V_{Imp} (%)
h_{crD50}	W	$f(T_{PF}) = 1.4E - 3T_{pf}^{0.51}$	0.79	166	< 0.001	0
h_{crD50}	h	$f(T_{PF}) = 1.2E - 3T_{pf}^{0.3}S^{-0.14}$	0.54	24	< 0.001	0
h_{crD50}	A_{xs}	$f(T_{PF}) = 7E - 7T_{pf}^{0.88}$	0.72	104	< 0.001	0
h_{Tree}	W	$f(T_{PF}) = 7E - 2T_{pf}^{0.31}$	0.81	161	< 0.001	0
h_{Tree}	h	$f(T_{PF}) = 3.9E - 1T_{pf}^{0.08}I^{0.94}$	0.38	11	< 0.001	22
h_{Tree}	A_{xs}	$f(T_{PF}) = 5E - 3T_{pf}^{0.47}I^{-0.34}$	0.77	65	< 0.001	7

Table 5.4: Reference equations of change ratios for W, A_{xs} , and W/h. Reference equations represent the best fitting functional relationship for the llow group of field sites for the NE region of Puerto Rico. The sample size (n), the coefficient of determination (\mathbb{R}^2), and the P-value (P_{val}) are reported as well.

Reference	Variable	Model	n	\mathbb{R}^2	P_{val}
h_{crD50}	W	$f(T_{PF}) = 4.1E - 4T_{pf}^{0.59}$	20	0.93	< 0.001
h_{crD50}	A_{xs}	$f(T_{PF}) = 6.3E - 7T_{pf}^{0.76}$	18	0.85	< 0.001
h_{crD50}	h	$f(T_{PF}) = 2.7E - 4T_{pf}^{0.43}$	15	0.71	< 0.001
h_{Tree}	W	$f(T_{PF}) = 1.2E - 1T_{pf}^{0.28}$	18	0.90	< 0.001
h_{Tree}	A_{xs}	$f(T_{PF}) = 5.9E - 2T_{pf}^{0.32}$	18	0.88	< 0.001
h_{Tree}	h	$f(T_{PF}) = 2.08E - 1T_{pf}^{0.23}$	16	0.83	< 0.001



Figure 5.1: Land use and precipitation (inset) map for the northeast region of Puerto Rico. Locations of field sites are represented by gray circles. Precipitation data is from the PRISM project and represents the 30 year average yearly precipitation in mm/yr. Land use and topography data are available from the US Geological Survey.



Figure 5.2: (a)-(f). Photographs of six different field sites displaying river name, drainage area (DA), average yearly precipitation (P), percentage of the basin covered in impervious cover (I), and width of channel at the surface of the water in the photograph (W). Photographs A, B, and C represent sites in the I_{low} group with increasing drainage area. Photographs D, E, and F represent sites in the I_{high} group with increasing drainage area.

Photographs A, D, E are tributaries of their respective labeled rivers.



Figure 5.3: (a) Example cross-section from the Rio Casa Blanca depicting reference depths, roughness measurement depth, and morphologic breaks. Black squares indicate surveyed points. (b) Relationship between depth at morphologic breaks and h_{crD50} . The solid red line indicates the best fitting regression line. Three sites were considered outliers, marked as diamonds, and excluded from the regression. Outliers were removed because morphologic breaks were obscured by large boulders. The dashed gray line represents the one to one line where h_{crD50} is exactly equal to the depth at the morphologic break



Figure 5.4: Unit flow duration curves of 15 min instantaneous discharge data for USGS stream gaging stations in NE Puerto Rico. Discharge data are from the years 2000-2009. Each discharge record is standardized by the drainage area for the USGS gaging station. Blue dashed lines represent I_{low} sites, solid black lines represent I_{med} sites, and red dotted lines represent I_{high} sites.



Figure 5.5: Range of grain size for the 16th, 50th, and 84th percentiles of the grain size distribution. Note that grain size is on a logarithmic scale. The sample size follows the percentage impervious in (). The median is marked by the dark line within the box. The upper and lower lines of the box represent the 3rd and 1st quartile, respectively. Dashed lines represent the maximum and minimum values of the data. Crosses represent data greater than 1.5 times the inner quartile range. The two insets represent the fraction of sand present in the grain size distribution and the fraction of the grain size distribution that is made up of anthropogenic debris.



Figure 5.6: Roughness values and average yearly total precipitation flux (T_{PF} on log scale). Blue circles, black crosses, and red X represent impervious cover groups I_{low} , I_{med} , and I_{high} , respectively. Manning's n values computed from Equation 5.1.



Figure 5.7: Ternary diagram representing exponents from hydraulic geometry data based on field measurements by the USGS. Exponents represent velocity(m), width (b), and depth (f). Blue circles represent I_{low} , black crosses represent I_{med} , and red X represent I_{high} sites.



Figure 5.8: Change ratios by reference depth, and amount of impervious cover. The sample size follows the percentage impervious cover. (a) Change ratios for stream width. (b) Change ratios for stream A_{xs} . (c) Change ratios for stream W/h ratios.

CHAPTER 6 : Summary and Conclusions

6.1. Summary

In this thesis I have attempted to illuminate one aspect of how climate interacts with mountainous landscapes, through bed load transport. I started by first asking the question of how a flood moves sediment, and from there I asked the simple follow-up question of how numerous floods move sediment (Chapters 2 and 3). Using the Mameyes River as a large-scale field experiment, I tracked the displacement of a plume of tagged cobbles at the annual and individual flood scale. Results from the individual floods showed that the most probably displacement for near-threshold floods is a single step, and that these displacements follow a thin-tailed distribution. To look at dynamics beyond a single flood, I needed to determine a way in which to quantify the variability within the hydrograph. Finding inspiration in particle-scale experiments (*Diplas et al.*, 2008), Chapter 2 developed a framework to quantify the flow variability by focusing on the integrated excess momentum within a flood, or the impulse. Furthermore, it was found that when the distribution of stress above the threshold of motion is thin-tailed, the impulse reduces to a dimensionless time. The characteristic behavior of the annual tracer displacement was found to scale linearly against dimensionless impulse (I_*) . These surveys also indicated that for numerous floods, the tracer particles remained in the regime of partial bed load transport.

Using the knowledge that the dimensionless impulse is akin to time in the Mameyes River, we found that tracer dispersion was superdiffusive. At first glance this result was not altogether obvious, as the expectation is subdiffusion for particles whose dispersion is dominated by long resting periods. However, rivers transport sediment in one direction, indicating that a population of sediment tracers undergoes an asymmetric random walk. Theory has demonstrated that power-law resting times lead to superdiffusion under the limiting case of strongly asymmetric transport. Using asymmetric random walk theory we were able to estimate that the distribution of particle waiting times approaches that of a heavy-tailed power-law. This scaling suggests that the particle waiting times exert a primary control on the downstream dispersal of sediment tracers. There is surprising agreement between these results and the experimental results of *Martin et al.* (2012). It is remarkable that results derived using pea gravel in a 2-m flume, and my results from a km-scale field experiment with cobbles, are the same. Recent theory suggests that the observed scaling for bed-load tracer dispersion is a generic consequence of burial and excavation of particles in a disordered system (*Martin*, 2013), suggesting that agreement between laboratory and field observations is not mere coincidence. Interestingly, field tracers also show good agreement with experiments on the sorting of particles by selective deposition. Field-deployed tracers quickly adjusted their sorting profile to match that of the stream bed. Another important result from this work is that particle displacements for two populations of tracers, in rivers with very different morphologies, can be collapsed onto a single linear function of dimensionless impulse by explicitly accounting for the threshold of motion and bed roughness in each stream. This implies that there may be a universal relationship between the average transport distance and the stress available to move particles.

Chapter 4 explored the implications of a dominant flood and equilibrium channel theory on the statistical scaling of the hydrograph for the Mameyes River. The Mameyes River was found to have a constant ratio of excess Shields stress across varying stream morphologies, ranging from a steep step-pool stream to the gravel-sand transition. The Mameyes River possesses a heavy-tailed distribution of discharge, but a thin-tailed distribution of shear velocity. That the Mameyes River exhibits both scaling behaviors demonstrates that heavytailed scaling of discharge does not indicate heavy tailed stresses. The shear velocity above the threshold of motion follows an exponential distribution, with a mean value that matches the bankfull flood. Furthermore, the dimensionless impulse is found to possess a unimodal peak that coincides with the bankfull flood, suggesting that the channel is adjusted to a characteristic impulse as well as stress. Using a large data compilation, the results from the Mameyes case study are observed to be general for gravel-bedded alluvial and bedrock channels. For all gages used in the data compilation, it was found that shear velocity scales as an exponential, and that the mean of the exponential function coincides with the bankfull flood. That the mean stress equals the bankfull stress indicates that one may be able to calculate the threshold stress from the hydrograph and an estimate of bankfull. To the authors knowledge these are the first observations linking the statistical distribution of the applied stress to channel form. The dimensionless impulse has a unimodal distribution with a peak that coincides with the bankfull flood, suggesting that channels are adjusted to the combination of a characteristic stress and duration. Taken together these observations indicate that all of the rivers analyzed are near threshold rivers adjusted to the bankfull flood. This reinforces and extends two central tenets of geomorphology – the principle of maximum geomorphic work, and equilibrium channel theory – and shows how climatic forcing may be simplified for treatments of long-term landscape evolution.

Chapter 5 explored the effects of impervious area on alluvial channel morphology in a humid tropical region. Field surveys of channel morphology at physically-based reference heights were compared across a gradient of land use, stretching from pristine tropical rain forests to high-density metropolitan areas. Despite large differences in land use between the surveyed streams, the differences in morphology are slight when drainage area and precipitation are accounted for. This was hypothesized to occur due to this region's channels being adjusted to frequent, sediment-mobilizing floods, and catchments that efficiently generate storm flow. For this region it was found that the impact of impervious surfaces is muted when compared to streams in more temperate climates.

Ultimately the evolution of landscapes occurs through the erosion, transport, and deposition of sediment. In mountainous regions this occurs through the transport of coarse sediment. A goal of this dissertation has been to provide a physical link between sediment transport dynamics at the particle level, and external forcing (climate, land use). Through the use of tracer particles this work has examined or inferred the detailed dynamics of coarse sediment transport at timescales ranging from seconds to years (7 orders of magnitude), and spatial scales of centimeters to kilometers (5 orders of magnitude). The resulting view is that coarse-grained rivers are adjusted to near threshold conditions, with bed load transport existing predominantly in the regime of partial transport. These observations hold even with heavy-tailed distributions of climatic forcing, as the threshold of motion filters extreme distributions in discharge into thin-tailed distributions of applied stress.

6.2. Implications and future prospects

This section highlights some of the broader implications, short falls, and future opportunities afforded by the work in the preceding chapters. The dimensionless impulse presents a potentially unifying framework within which to view the dynamics of sediment tracers. The linear scaling between the mean displacement and the dimensionless impulse indicates that the mean transport dynamics of bed load are governed by momentum conservation, as determined from two end-member bed morphologies. Data suggest that there may be a universal scaling among bed load tracers, allowing one to predict the displacement of a population of coarse particles from the hydrograph. These observations are important for river restoration as they provide a physical framework in which to predict the behavior of sediment introduced, for example, during a gravel augmentation project below a dam (Schmidt and Wilcock, 2008) or by a landslide. While the mean scaling is determined by the fluid momentum (an allogenic signal), it is the particle waiting times (autogenic) that control the particle dispersion; and these are controlled by bed elevation fluctuations due to particle-particle interactions. The superdiffusive scaling may limit the application of the momentum-conservation approach at long timescales because the signal of mean displacement (climate) may be swamped by the rapid growth in variance (e.g., Jerolmack and Paola, 2010; Jerolmack, 2011).

Viewing landscapes and sediment transport through the lens of the threshold of motion has proven to be a very beneficial approach in this dissertation. In general this is an approach that is, surprisingly, under utilized in fluvial geomorphology. Perhaps it is due to the difficulty in measuring the threshold of motion, which lacks a standardized method of determination; different methodologies give very different values (*Wilcock*, 1988). It is also not currently known how dynamic the threshold of motion is, in that some observations suggest it can change from flood to flood (*Turowski et al.*, 2011). That one may be able to determine the threshold of motion from the hydrograph and an estimate of the bankfull stress is a remarkable result (chapter 4), though it requires further exploration and validation. An immediate avenue to explore is the timescales over which the channel geometry adjusts, in order to determine the duration of a hydrograph that is necessary to acquire a representative estimate of the threshold of motion. Following the analysis of *Jerolmack and* Mohrig (2007), one hypothesis is that this timescale is equal to the time necessary to transport a volume of sediment equal to the channel dimensions. The work in this dissertation shows that the threshold of motion is one of the most essential parameters to understanding fluvial dynamics in gravel-bedded rivers. The following outlines a simple example where this approach could yield promising results. For aquatic organisms living in rivers, the stream bed is a fundamental part of the habitat. Knowledge of the threshold of motion represents a way to calculate the timescales over which this habitat becomes rearranged. It has been shown that this timescale of bed rearrangement can fundamentally limit the types of ecological communities and food webs within a river (*Power et al.*, 2008).

Interestingly, bed load transport was only one of numerous systems in which Wolman and Miller (1960) tested the concept of maximum geomorphic work. Therefore it is possible that the thin-tailed distribution of stresses above the threshold (chapter 4) holds in other natural sediment transport systems such as: aeolian dune fields, sandy rivers, suspended sediment load, and the profile of a gravel or sandy beach (Wolman and Miller, 1960). This thin-tailed scaling may be a general feature of natural systems that are adjusted to a maximum of geomorphic work; there is preliminary support for the generality of near-threshold transport and thin-tailed stresses in aeolian dune fields (Jerolmack and Brzinski III, 2010).

The impulse framework (developed in chapter 2) is not without its short comings, as it is exquisitely sensitive to the determination of the threshold of motion. Furthermore, the impulse framework only accounts for hydrodynamics, in a problem that may be strongly
influenced by granular effects (*Frey and Church*, 2011). The linear scaling of the dimensionless impulse and the mean tracer displacement suggests that variability in hydrodynamic forcing may not be that important in determining sediment transport statistics; however, observations of the effects of flood sequence, magnitude, and duration suggest that at short timescales the variability in the flow exerts some influence (*Hsu et al.*, 2011; *Turowski et al.*, 2011; *Yager et al.*, 2012). In a sense these, effects impart a memory to the system that the linear scaling does not account for; but this memory seems to average out at the annual scale, preserving the linear scaling. The mechanisms behind this mesoscale averaging are not well known. It is currently an open question as to whether a short duration, high-magnitude flood produces different granular dynamics from a long-duration, low-magnitude flood, if each have the same dimensionless impulse. Following the completion of this dissertation, I plan to turn my attention to unraveling the interactions between flood shape, sequences of floods, and granular phenomena. I intend to focus on how and why a simple fluid momentum approach, such as the dimensionless impulse, might break down in predicting sediment transport.

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