BEDLOAD TRANSPORT, FLOW HYDRAULICS, AND MACROROUGHNESS IN STEEP MOUNTAIN STREAMS

A thesis submitted to
ETH ZURICH
for the degree of
DOCTOR OF SCIENCES
(Dr. sc. ETH Zurich)

presented by

JOHANNES MARTIN SCHNEIDER

Dipl. Hyd., Albert-Ludwigs-University Freiburg i. Br.
born September 14, 1982
citizen of Germany

accepted on the recommendation of

Prof. Dr. J.W. Kirchner (examiner)
Dr. D. Rickenmann (co-examiner)
Dr. J.M. Turowski (co-examiner)
Prof. Dr. S.N. Lane (co-examiner)

2015
# TABLE OF CONTENTS

**Summary** .................................................................................................................. III

**Zusammenfassung** ....................................................................................................... V

**Chapter I** .................................................................................................................. 1
  - Background ............................................................................................................ 1
  - Objectives ............................................................................................................. 4
  - Structure of the thesis ......................................................................................... 7
  - References ............................................................................................................ 7

**Chapter II** .................................................................................................................. 13
  - Stream bed roughness and flow velocity in a steep mountain channel ............. 13

**Chapter III** .................................................................................................................. 43
  - Applicability of bedload transport models for mixed size sediments in steep streams considering macro-roughness .......... 43

**Chapter IV** .................................................................................................................. 79
  - Scaling relationships between bedload volumes, transport distances and stream power in steep mountain channels .......... 79

**Chapter V** .................................................................................................................. 105
  - Conclusions ......................................................................................................... 105
  - References ............................................................................................................ 108

**Appendix** ................................................................................................................... 109
  - Supplementary material: Study I (Chapter II) ............................................... 109
  - Supplementary material: Study II (Chapter III) ............................................. 117
  - Supplementary material: Study III (Chapter IV) ........................................... 129

**Publication List** ........................................................................................................ 137

**Dank** ......................................................................................................................... 139
SUMMARY

Bedload transport in steep mountain streams is a key element in the evolution of the landscape, in hydraulic engineering, and natural hazard prediction. Thus a better process understanding of bedload transport and its driving forces is essential for accurate prediction of bedload transport necessary for the dimensioning of hydraulic structures, for morphodynamic predictions and for minimizing potential damages. However, data for model development and validation are scarce in steep mountain streams compared to lowland rivers and the transport processes are still not fully understood. In addition, the conditions driving bedload transport in steep streams are typically more complex compared to lowland rivers. Increased bed heterogeneity at steeper bed gradients typically results in increased flow resistance and limited sediment availability, and these factors greatly affect bedload transport rates. The present thesis is concerned with bedload transport and its driving forces in steep mountain streams, i.e. (i) the influence of streambed roughness on flow velocity and thus on the flow energy available for bedload transport, (ii) how a limited energy available for bedload transport can be considered in bedload transport equations and (iii) how bulk bedload transport is related to the transport distances of individual moving particles.

The first study focused on the interactions between streambed, roughness and flow hydraulics over a wide range of bed gradients and related bed roughness characteristics. It is based on high-quality field data from a steep mountain stream, the Riedbach (canton Valais, Switzerland). The Riedbach is characterized by almost flume-like boundary conditions. Bed gradient increases along the 1-km study reach by about one order of magnitude (S=3-41%), with a corresponding increase in streambed roughness, while flow discharge and width remain approximately constant. The streambed roughness along the study reach was characterized by measures derived from terrestrial laser scanning point clouds and characteristic grain sizes. Flow velocity was derived from dye tracer experiments. Flow resistance approaches were used to relate the measured streambed roughness to flow velocity. Flow resistance behavior across this large range of steep slopes agrees with patterns established in previous studies, mainly for lower gradient reaches, irrespective of the roughness measures that are used. Furthermore, empirical critical shear stress approaches were linked to the variable power equation for flow resistance to investigate the change of bed roughness, i.e. the characteristic grain size D84, with channel slope. The predicted increase in D84 with increasing channel slope was in good agreement with the field observations.

The second study investigated how flow energy available for bedload transport is reduced at steep gradients and how this affects bedload transport predictions. Bedload transport models for mixed size sediments and models based on a median grain size were evaluated using a large field data compilation of fractional bedload transport rates, including streams with bed gradients ranging from 0.05% to 11%. Using existing bedload transport relations or modifying them, suitable values were derived for the reference shear stress (a measure similar to the critical shear stress describing the initiation of motion), using either the total boundary shear stress, or the reduced (termed “effective”) shear stress that accounts for flow resistance.
due to macro-roughness. Both the reference shear stress and the bedload transport relations strongly depend on how the acting shear stress is calculated. However, considering the reductions in flow energy available for bedload transport due to macro-roughness in steep mountain streams, allows transport equations developed on flume data for lower gradient channels to be applied for steep streams.

The third study considers bulk bedload transport volumes during storm events, calculated as the product of the width and depth of bed scour and the (mean) transport distances of individual grains. Bed load transport rates were measured in two steep mountain streams, in the Erlenbach (Switzerland) and the Rio Cordon (Italy). Individual gravels were traced using magnetic and radio frequency identification tags. Tracer transport distances and bed load volumes exhibited approximate power law scaling with peak stream power of individual hydrologic events. Bed load volumes scaled much more steeply with peak stream power than tracer transport distances did, and bed load volumes scaled as roughly the third power of transport distances. These observations imply that large bed load transport events become large primarily by scouring the bed deeper and wider, and only secondarily by transporting the mobilized sediment farther. Using the sediment continuity equation, it was possible to estimate the mean effective thickness of the actively transported layer, averaged over the entire channel width and the duration of individual flow events. The derived active layer thickness, ranging up to 0.57m in the Erlenbach, is broadly consistent with independent measurements. Thus, the active layer concept provides a useful simplified characterization of streambed response to hydraulic forcing during bedload transport.

Each of the studies presented within this thesis is based on rarely available, high quality field data on bedload transport and its driving forces in steep mountain streams. The findings contribute to a better understanding of transport processes and can be used for prediction of flow velocity, hydraulic stress and bedload transport.
ZUSAMMENFASSUNG


Die zweite Studie untersucht, wie sich die für den Geschiebetransport verfügbare Strömungsernergie, in Abhängigkeit der Gerinnereinigung und der Mindestbelastung für den Bewegungsbeginn bei grossem Gerinnegefälle reduziert. Basierend auf einem umfassenden Geschiebetransportdatensatz (33 Flüsse, 0.01-11% Gerinneneigung) wurde die
Referenzschubspannung (ein Mass ähnlich zur kritischen Schubspannung für den Bewegungsbeginn) sowie Geschiebetransportgleichungen abgeleitet. Dabei wurde wahlweise eine voll wirkende („totale“) Sohlschubspannung und eine reduzierte („effektive“) Schubspannung angenommen, wobei in der reduzierten Spannung Fließwiderstände durch Makro-Rauigkeiten berücksichtigt werden. Sowohl die abgeleitete Referenzschubspannung als auch die Geschiebetransportgleichungen sind stark von der Wahl der Berechnungsmethode für die wirkende Sohlschubspannung abhängig. Transportgleichungen, die für niedrigere Gefällsbereiche entwickelt wurden, lassen sich in steilen Gebirgsbächen anwenden wenn Energieverluste durch Gerinnerauigkeit berücksichtigt werden.


CHAPTER I

BACKGROUND

Bedload transport in steep headwater streams, throughout this study defined by bed gradients larger than 2%, is a major agent in fluvial geomorphology [Schumm, 1977], as it controls the erosion of sediments and transfer to lower-gradient rivers. Bedload transport refers to particles that move by rolling, sliding or saltation along the streambed (in contrast to suspended sediments) and is a function of sediment sizes and load, bed gradient and flow discharge rates [e.g., Charlton, 2007]. The available and potentially transported sediments regulate channel stability [e.g., Church and Zimmermann, 2007], aquatic habitats [e.g., Buffington et al., 2004; Chapman, 1988; Goode et al., 2012], groundwater exchange [e.g., Bencala, 2011; Lambs, 2004], hydraulic conditions [e.g., Powell, 2014] and bedrock erosion [e.g., Turowski, 2009; Turowski and Rickenmann, 2009]. Furthermore, bedload transport in mountain streams has to be considered in a wide range of river engineering applications, i.e. for hydropower plants, sediment reservoirs, water intakes or dams [e.g., Yang, 1996]. Also, bedload transport connected to large flood events can cause enormous damages in mountain regions [Badoux et al., 2014].

For an accurate prediction of bedload transport rates, the design of constructions and natural hazard prediction, a better process understanding of bedload transport and its driving forces is essential. However, despite decades of research, the processes governing bedload transport in steep streams are not fully understood. Furthermore, the transferability of the knowledge and equations for bedload transport in low-gradient rivers to headwater streams is limited because of the changing boundary conditions affecting bedload transport.

In this thesis it is shown how increases in stream bed heterogeneity (i.e. coarsening and widening grain size distributions) with increased bed gradients impact stream dynamics. Changing grain size distributions with increasing bed gradients have significant impacts on the bedload transport behavior as they (i) control flow resistance and thus the flow energy available for bedload transport, (ii) control bed stability and thus the energy needed for initiation of motion and finally (iii) control the sediment availability for transport. Within the present thesis, I concentrate on effects (i) and (ii), while effect (iii) is only touched on. The three effects are introduced in more detail in the following sections.

(i) Flow resistance

Moving upstream from large, low-gradient rivers towards steep headwater streams, one can readily observe how the grain size distribution coarsens [e.g., Gomez et al., 2001; Knighton, 1980; Rice and Church, 1998], connected with channel bed characteristics changing from pool-riffle to plane-bed, step-pool and cascade morphologies [Montgomery and Buffington, 1997]. Furthermore, the abundance of immobile boulders increases [e.g., Nitsche et al., 2011; Yager et al., 2007], the channel width becomes more variable, and the number of bedrock constrictions and the frequency of large woody debris may increase [Wohl, 2000]. These factors are generally described as macro (or form) roughness elements
and contribute to an increased flow resistance at steeper gradients and thus reduced proportions of the flow energy available for bedload transport [e.g., Nitsche et al., 2011; Rickenmann and Recking, 2011].

Quantifying flow resistance and flow velocity is most commonly based on flow resistance equations as e.g. the Chezy, the Manning and the Darcy-Weisbach relations. These equations are based on flow resistance related coefficients (i.e. the Chezy coefficient $C$, the Manning coefficient $n$ or the Darcy-Weisbach friction factor $f$). Many studies have attempted to link these coefficients with bed characteristics and flow conditions, i.e. scaling flow depth with a roughness height, and in many cases the characteristic grain size $D_{S4}$ was selected [Aberle and Smart, 2003; Comiti et al., 2009; Comiti et al., 2007; Ferguson, 2007; Zimmermann, 2010], for deriving generally valid equations [e.g., Limerinos, 1970; Strickler, 1923] (recent comprehensive reviews on flow resistance prediction based on these roughness coefficients can be found e.g. in Powell [2014], Rickenmann and Recking [2011] and Yochum et al. [2012]).

As an alternative to scaling flow depth by a roughness measure, dimensionless hydraulic geometry relations were introduced which scale flow velocity and discharge using a roughness height, such as $D_{S4}$ [Ferguson, 2007], and additionally the channel bed slope [Nitsche et al., 2012; Rickenmann and Recking, 2011]. These quantitative methods provide generally very good estimates of flow velocities, especially in comparison with the previously introduced concept using scaled flow depth and e.g. the Darcy-Weisbach friction factor [e.g., Comiti et al., 2009; Nitsche et al., 2012; Rickenmann and Recking, 2011]. This is at least partly because accurate flow depth estimates are typically more difficult to obtain in steep streams with irregular bed forms [Rickenmann and Recking, 2011].

In order to account for flow resistance due to form (macro) roughness, additional approaches were developed going back to flow resistance partitioning into grain and form resistance as introduced by Meyer-Peter and Mueller [1948] [e.g., Carson, 1987; Carson and Griffiths, 1987; Gomez and Church, 1989; Millar and Quick, 1994; Millar, 1999; Palt, 2001; Parker and Peterson, 1980; Rickenmann and Recking, 2011; Rickenmann et al., 2006]. Energy losses due to form roughness in steep streams are typically described by empirical equations based on roughness measures such as the characteristic grain size $D_{S4}$ [Rickenmann and Recking, 2011] or $D_{S3}$ [Wilcock et al., 2009], but they were also determined as a function of gravel bars and pool-riffle sequences [Millar and Quick, 1994; Millar, 1999], boulder concentration [Pagliara and Chiavaccini, 2006; Whittaker et al., 1988], boulder protrusion [Yager et al., 2007] or step height and spacing [Egashira and Ashida, 1991].

(ii) Initiation of motion

A second effect of a coarsened grain size distribution at steep bed gradient is that bed stability is increased due to interlocking of bed particles [Church et al., 1998], enhancing the forces needed for particle entrainment. Furthermore, variable friction angles, grain emergence, increased turbulence and flow aerations (lower water density) may increase total drag and contribute to the (observed) increased shear stress for the initiation of motion of individual particles [Lamb et al., 2008]. A commonly used parameter describing the critical conditions for initiation of particle motion in bedload transport calculations [e.g., Bathurst,
2013; Wilcock, 1988] is the ‘critical’ shear stress [Shields, 1936] defining the maximum particle size that can be moved at a given flow. An increase in the critical shear stress with increasing channel slope has been shown by several authors [e.g., Bunte et al., 2013; Camenen, 2012; Ferguson, 2012; Lamb et al., 2008; Prancevic et al., 2014; Recking, 2009; Shvidchenko et al., 2001]. Other studies have identified relationships for the reference shear stress (which is used as a parameter equivalent to the critical shear stress) accounting for the characteristics of the grain size distributions more directly, i.e. as a function of sand content [Wilcock and Crowe, 2003], the geometric standard deviation of the bed surface grain size distribution [Gaeuman et al., 2009] or the degree of armouring [Efthymiou, 2012].

Furthermore, the coarsening and widening of the streambed grain size distributions with increasing bed gradients potentially affect the critical conditions for the onset of motion of individual particle size classes. Clearly, particle mobility is related to absolute particle size, and larger, heavier particles are harder to mobilize than smaller, lighter particles [Buffington and Montgomery, 1997; Shields, 1936]. However, in a bed with mixed grain sizes, not only the absolute grain size but also the relative grain size affects particle mobility. Smaller particles are hidden behind larger pieces of gravel, and are therefore protected from the flow, while larger particles are more exposed to the flow [Egiazaroff, 1965; Einstein, 1950; Parker, 2008; Parker and Klingeman, 1982; Parker et al., 1982; Wiberg and Smith, 1987]. These effects are known as hiding and exposure, and they can modulate the relative mobility of different grain size fractions. If hiding and exposure effects exactly cancel out weight effects, the probability of entrainment for each grain becomes independent of grain size at a given hydraulic stress, resulting in a condition termed ‘equal mobility’. The initiation of motion of individual particle sizes classes related to relative grain-size effects are taken into account by fractional transport models which typically include an empirically derived hiding function to describe the mobility of particles that are fine or coarse in relation to the median bed size [e.g., Andrews et al., 1987; Ashida and Michiue, 1973; Parker, 1990; Parker and Klingeman, 1982; Parker et al., 1982; Powell et al., 2001; Wilcock and Crowe, 2003].

(iii) Sediment supply

Sediment supply is related to the available sediment that can be entrained and transported by flows larger than the transport threshold and refers to sediments either stored in the channel bed or supplied from upstream, from the banks or hillslopes. The sediment supply is closely related to the sediment sizes and topographic conditions, which typically change in terms of decreased sediment availabilities at steep streams compared to lowland rivers. For example, sediment availability is decreased at steep slopes because the proportions with relatively stable patches is increased [Nelson et al., 2009; Yager et al., 2012a] due to topographic controls as bar morphology [Laronne and Duncan, 1992; Lisle and Hilton, 1999] or accumulations of immobile boulders [e.g., Garcia, 1999; Laronne, 2001]. Bed armouring often is due to the development of a coarse immobile surface layer, which protects a finer subsurface layer from the flow [Bathurst, 2007]. Furthermore, the alluvial cover in steep stream reaches is usually thinner, the percentage of bedrock is increased and sediment entries from upstream are smaller compared to lower gradient reaches.
Introduction

These processes related to the heterogeneity of the stream bed have important effects on bedload transport behavior. On the one hand the transport capacity is considerably reduced at steep slopes compared to lower gradient rivers due to increased flow resistance (i) and an increased demand of energy for initiation of motion due to structural bed stability (ii). On the other hand sediment availability might be limited at steep slopes (iii). If these effects, typical for steep streams, are not considered in bedload transport equations (typically developed for low-gradient rivers) and calculations, bedload transport might be overestimated by orders of magnitude [Bathurst et al., 1987; Chiari and Rickenmann, 2011; Lenzi et al., 1999; Rickenmann, 2001; 2012; Yager et al., 2007; Yager et al., 2012b].

OBJECTIVES

Within the studies presented in this thesis, it is aimed to contribute to a better process understanding of bedload transport in steep mountain streams regarding the following aspects:

- How is flow velocity related to streambed characteristics such as grain size distributions and bed roughness heights at increasing bed gradients (Study I)?
- How do critical conditions for particle entrainment change with bed gradient and how does the choice of estimating the transport capacity affect the shape of fractional and total transport equations (Study II)?
- How is bulk bedload transport related to the movement of individual particle transport distances and the sediment available for transport (Study III)?

According to these objectives, I concentrate on the effects (i) flow resistance and (ii) initiation of motion on bedload transport, while the effect (iii) sediment availability is touched upon but not studied in detail.

Data compilation

For the validation and further development of existing flow resistance and bedload transport approaches, high quality field data are essential. Because field data on bed, flow and bedload transport characteristics in steep mountain streams are sparse, the first aim within this thesis was to collect high quality data on these characteristics in two steep mountain streams, namely the Riedbach (canton Valais) and the Erlenbach (canton Schwyz). The following two paragraphs will give a short overview on the field sites and the main field measurements conducted during the study.

The Riedbach is a glacial fed mountain stream characterized by bed gradients ranging from 3% to 41%. According to the bed gradient, also the channel bed morphology, the surface grain size distributions and the bed roughness change considerably. However, flow discharge and stream width only marginally change along the 1 km long study reach due to the glacial runoff regime and given morphological conditions, respectively. This setup of the Riedbach provided almost flume-like conditions for investigations on bed, flow and bedload transport interactions. The channel bed was characterized in detail by measurements of grain size distributions, boulder density and protrusion, longitudinal and cross-sectional profiles. In
addition, a terrestrial laserscan was performed to derive a detailed channel bed topography and different kinds of roughness measures for characteristic stream reaches, namely the glacier forefield, the transitional domain and the steep reaches. Flow characteristics were determined from 46 dye tracer experiments, during which a dye tracer was injected on the glacier forefield or upstream of the steep reach. Using ten fluorometers the dye tracer break through curves along the longitudinal profile was measured in order to determine reach averaged flow velocities at changing discharges and increasing bed gradients. Finally bedload transport was measured on the glacier forefield using portable bedload traps [see also Bunte et al., 2013; Schmid, 2011] and at the downstream end of the steep reach using a calibrated geophone system [Schneider et al., 2013]. In addition stationary RFID (radio frequency identification) antennas were developed [Schneider et al., 2010] to determine mean residence times of individual tracer stones (data and results derived from the Riedbach geophone and RFID systems are not part of the thesis papers).

The Erlenbach is a steep mountain stream (mean gradient: 17%) draining an area of 0.7 km² in the Swiss pre-Alps. The runoff regime is nivo-pluvial (snow and rain dominated) with the largest transport events caused by extreme summer rainstorms. The Erlenbach field site provides continuous data on flow discharge as well as bedload transport rates based on a gauging station and a geophone system, respectively [e.g., Rickenmann and McArdell, 2007; Rickenmann et al., 2012; Turowski and Rickenmann, 2011]. In addition, an automatic moving bedload basket system was installed in 2009, which provides fractional bedload transport rates together with discharge measurements [Rickenmann et al., 2012]. In 2009 and 2010, particle tracking studies were conducted, by equipping natural particles with RFID transponders and following their pathways after bedload transport events [Hegglin, 2011; Meier, 2009].

Study I

Stream bed surface characteristics, flow velocity and bedload transport are closely related to each other. However, currently there are only few field data available from very steep slopes justifying or contradicting existing flow resistance approaches relating flow velocity to streambed roughness. Within this study we quantified these interactions and their change with increasing bed gradients, based on the detailed TLS derived bed topography measurements and dye tracer flow velocity measurements. The questions of the study were:

- How do grain size distributions and bed roughness change with increasing bed gradients?
- How does flow velocity change with increasing bed gradients?
- Which bed roughness measure is best suited to flow velocity prediction (assuming a fixed, stationary bed roughness)?
- Can flow resistance equations be applied to almost unmeasured, very steep bed slopes?
- How does the channel bed adjust with respect to flow velocity and bedload transport at given stream gradient
Introduction

This study is presented in Chapter II and has been submitted as a research article to Water Resources Research: Schneider, J. M., D. Rickenmann, J. M. Turowski and J. W. Kirchner (2015): Stream bed roughness and flow velocity in a steep mountain channel.

Study II

Accounting for a reduced transport capacity at steep slopes is commonly based on reducing the energy available for bedload transport or increasing the critical conditions for initiation of motion because the individual contributions of two factors are difficult to quantify from field data. However, it has not been shown in detail how the choice for one or the other approach affects bedload transport prediction. Furthermore, transport equations were developed to calculate bedload transport for each grain size fraction separately but to date these approaches were only rarely tested with field data of steep mountain streams. Based on fractional bedload transport data from 10 US streams, the Riedbach, the Erlenbach and additional data available in literature, answers regarding the following questions are provided:

- What is the critical (reference) shear stress for a given particle size class if derived from either total boundary or a reduced (taking flow resistance due to macro-roughness into account) shear stress?
- How can the critical shear stress for the mean grain size or for individual particle size classes be explained by stream or flow characteristics?
- How does the choice of reducing the acting shear stress affect the shape of a bedload transport equation and thus bedload transport prediction?
- Is the shape of fractional transport equations similar to the shape of a total transport equation?
- Does fractional transport calculation provide advantages in comparison to total transport calculation (besides providing fractional transport rates)?

This study is presented in Chapter III and has been submitted as a research article to Water Resources Research: Schneider, J. M., D. Rickenmann, J. M. Turowski, Bunte, K., and J. W. Kirchner (2014): Applicability of bedload transport models for mixed size sediments in steep streams considering macro-roughness.

Study III

Bedload transport is commonly determined based on empirically derived bedload transport equations. Another approach to estimate bedload transport is based on the notion that the bulk bedload transport is controlled by individually moving particles. However only very limited data are available on bulk bedload transport and transport distances of individual transport events. Based on the data from two mountain streams, the Erlenbach and the Rio Cordon (Italy), answers to the following questions are provided:

- How are bedload volumes and transport distances related to peak stream power and the cumulative stream energy of individual hydrologic events?
Chapter I

- How are bedload volumes related to mean transport distances?
- What is the primary contributor for increasing bedload volumes at increasing transport event magnitudes? Increasing transport distances or increasing thickness of the mobile particle layer, termed active layer depth?

This study is presented in Chapter IV and has been published in the Journal of Geophysical Research – Earth Surface: Schneider, J. M., J. M. Turowski, D. Rickenmann, R. Hegglín, S. Arrigo, L. Mao and J. W. Kirchner (2014): Scaling relationships between bedload volumes, transport distances and stream power in steep mountain channels.

STRUCTURE OF THE THESIS

The thesis consists of five chapters. In Chapter I the introduction is given, the three studies are presented in Chapters II-IV. The conclusion of this work is summarized in Chapter V. The supplementary materials (related to the publications in Chapters II-IV) as well as a publication list of additional publications, which are not part of the thesis, are attached at the end of the thesis.

REFERENCES

Ashida, K., and M. Michiue (1973), Study on bed load transport rate in open channel flows paper presented at International Symposium on River Mechanics IAHR, Bangkok, Thailand
Buffington, J. M., and D. R. Montgomery (1997), A systematic analysis of eight decades of
Introduction


Ferguson, R. I. (2012), River channel slope, flow resistance, and gravel entrainment
Chapter I


Laronne, J. B., and M. J. Duncan (1992), Bedload transport paths and gravel bar formations, 177-202 pp., John Wiley & Sons Ltd, Chichester.


Introduction


Chapter I

VAW-ETH Zürich, WSL Birmensdorf.


Shields, A. (1936), Anwendung der Ähnlichkeitsmechanik und der Turbulenzforschung auf die Geschiebewegung, Triltsch & Huther.


Wilcock, P. R., J. Pitlick, and Y. Cui (2009), Sediment transport primer: estimating bed-material transport in gravel-bed rivers, US Department of Agriculture, Forest Service, Rocky Mountain Research Station.


Introduction


CHAPTER II

Stream bed roughness and flow velocity in a steep mountain channel

Johannes M. Schneider, Dieter Rickenmann, Jens M. Turowski, James W. Kirchner

ABSTRACT - Understanding how channel bed morphology affects flow conditions (and vice versa) is important for a wide range of fluvial processes and practical applications. We investigated interactions between bed roughness and flow velocity in a steep, glacier-fed mountain stream (Riedbach, Ct. Valais, Switzerland) with almost flume-like boundary conditions. Bed gradient increases along the 1-km study reach by roughly one order of magnitude ($S=3\text{-}41\%$), with a corresponding increase in streambed roughness, while flow discharge and width remain approximately constant due to the glacial runoff regime. Streambed roughness was characterized by semi-variograms and standard deviations of point clouds derived from terrestrial laser scanning. Reach-averaged flow velocity was derived from dye tracer breakthrough curves measured by 10 fluorometers installed along the channel. Commonly used flow resistance approaches (Darcy-Weisbach relation, dimensionless hydraulic geometry) were used to relate the measured bulk velocity to bed characteristics. As a roughness measure, $D_{84}$ yielded comparable results to more laborious measures derived from point clouds. Flow resistance behavior across this large range of steep slopes agreed with patterns established in previous studies for both lower-gradient and steep reaches, irrespective of which roughness measures were used. We linked empirical critical shear stress approaches to the variable power equation for flow resistance to investigate the change of bed roughness with channel slope. The predicted increase in $D_{84}$ with increasing channel slope was in good agreement with field observations.

1. Introduction

Flow velocity is an essential determinant of many fluvial processes and properties, including flood routing, transport of nutrients, pollutants, and sediments, and aquatic habitat quality. However, despite decades of research it is not fully understood how flow velocity is controlled in gravel-bed streams, especially in steep mountain channels where the bed morphology is typically complex and rough. As bed gradients steepen, flow resistance typically rises due to the increasing proportion of coarse roughness elements such as immobile boulders, bedrock constrictions or large woody debris. Furthermore, flow resistance strongly increases with decreasing relative submergence, i.e. the ratio of flow depth to a characteristic roughness size [Bathurst, 1985; Ferguson, 2010; Lee and Ferguson, 2002; Limerinos, 1970; Reid and Hickin, 2008; Wilcox and Wohl, 2006].

Grain, form and spill resistance are all important in steep streams. Whereas grain resistance is related to the skin friction and form drag on individual grains on the stream bed surface [Einstein and Barbarossa, 1952], form resistance is related to the pressure drag on irregular
Stream bed roughness and flow velocity

Bed surfaces [Leopold et al., 1995]. Spill resistance in step-pool or cascade streams [Montgomery and Buffington, 1997] results from turbulence when supercritical flow decelerates as fast flows meet slower-moving water [Leopold et al., 1995; Leopold et al., 1960; Wilcox et al., 2006]. Spill resistance is typically dominant in step-pool or cascade mountain streams [Abrahams et al., 1995; Comiti et al., 2009; Curran and Wohl, 2003; David, 2011; Wilcox et al., 2006; Zimmermann, 2010] and generally decreases with increasing flow stage [David, 2011; Wilcox et al., 2006].

Flow resistance relations may be used to predict flow velocity in steep streams where no direct measurements are available. Flow resistance calculations are typically based on the Manning or Darcy-Weisbach equations,

\[ v = \frac{S^{1/2}d^{2/3}}{n} = \sqrt{\frac{8gdS}{f_{tot}}} \]  

(1)

where \( n \) is the Manning coefficient [L^{-1/3}s], \( f_{tot} \) is the Darcy-Weisbach friction factor [dimensionless], \( S \) is the energy slope (in this study \( S \) is approximated by the channel bed slope), \( d \) is the flow depth (in this study \( d \) is approximated by the hydraulic radius, consistent with the conventional practice in narrow channels [Rickenmann and Recking, 2011]) and \( g \) is gravitational acceleration [ms^{-2}]. Many studies have identified relations between measures of roughness height \( R \) (e.g., \( D_{84} \)) and bed or flow characteristics; comprehensive overviews of flow velocity predictions based on \( n \) or \( f_{tot} \) can be found in Powell [2014], Rickenmann and Recking [2011], and Yochum et al. [2012]. In particular, low submergence of roughness elements (i.e. small relative flow depth \( d/R \)), typical for steep streams, was outlined to be an important agent for flow resistance [e.g., Aberle and Smart, 2003; David et al., 2010; Ferguson, 2010; Recking et al., 2008; Wohl and Merritt, 2008; Wohl et al., 1997; Yochum et al., 2012].

Alternatively, hydraulic geometry relations have been scaled and non-dimensionalized using a roughness height \( R \) [Ferguson, 2007] and channel bed slope [Nitsche et al., 2012; Rickenmann and Recking, 2011] to account for the increased influence of macro-roughness elements at steep slopes on flow velocity (Equations 2 and 3).

\[ v^{**} = \frac{v}{\sqrt{gSR}} \]  

(2)

\[ q^{**} = \frac{q}{\sqrt{gSR^2}} \]  

(3)

Here \( v^{**} \) is the dimensionless velocity and \( q^{**} \) the dimensionless discharge. These quantitative methods generally provide better estimates of (dimensional) flow velocities in steep streams, than approaches using scaled flow depth and \( f_{tot} \) do [e.g., Comiti et al., 2009; Nitsche et al., 2012; Rickenmann and Recking, 2011; Zimmermann, 2010]. One reason for this finding is the difficulty of measuring or defining representative flow depths in steep streams with irregular bed forms [Rickenmann and Recking, 2011].

When relating flow stage, flow velocity, discharge or resistance to bed characteristics, the \( D_{84} \) of the streambed surface layer is often selected as dominant roughness height \( R \) [Aberle and Smart, 2003; Comiti et al., 2009; Comiti et al., 2007; Ferguson, 2007; Zimmermann, 2010]. Other roughness measures to characterize flow resistance were also proposed,
including boulder concentration [Pagliara and Chiavaccini, 2006; Whittaker et al., 1988], boulder protrusion [Yager et al., 2007] or step height and spacing [Egashira and Ashida, 1991]. It is still an open question which roughness measure is most representative for flow resistance in steep and rough streams, considering that natural beds are composed of heterogeneously sized grains which are non-uniformly spaced and protrude into the flow to varying extents [Kirchner et al., 1990]. Field estimates of a characteristic grain size such as $D_{84}$ are often associated with large uncertainties due to operational bias [Marcus et al., 1995; Wohl et al., 1996], limited sample size [Church et al., 1987; Milan et al., 1999], incompatibility between sampling methods [Diplas and Sutherland, 1988; Fracarollo and Marion, 1995] and spatially heterogeneous grain size distributions [Buffington and Montgomery, 1999; Crowder and Diplas, 1997]. With the further development of photogrammetry and laser scanning technologies, there are new possibilities for obtaining detailed topographic information and streambed roughness measures, including the standard deviation, semi-variance, skewness or kurtosis of the detrended bed surface elevations [e.g. Heritage and Milan, 2009; Rychkov et al., 2012; Smart et al., 2002; Yochum et al., 2012].

Previously introduced approaches often consider bed roughness measured at a given moment as a constant parameter over time. Considering a stream as a self-organized system with feedback mechanisms between bed morphology, hydraulics and sediment transport, it may be assumed that bed roughness adjusts to bed gradient and flow conditions, reflecting the dominant hydraulic stresses that were responsible for the formation of the streambed. It is commonly argued that the bed adjusts to maximize flow resistance [e.g. Davies and Sutherland, 1980; Davies and Sutherland, 1983] because maximum flow resistance is connected with maximum bed stability [Abrahams et al., 1995]. Bed stability and bed adjustment are closely related to the grain sizes that can be entrained under specified bed gradients and flow conditions. A commonly used parameter that describes critical conditions of particle entrainment is the critical shear stress $\tau^*_c$, which incorporates in its dimensionless form a critical flow depth $d_c$, the bed gradient $S$ and the related grain size $D_s$ (Equation 4),

$$\tau^*_c = \frac{d_c S}{(s-1)D_s} \quad (4)$$

where $s$ is the relative density of the sediment ($s \approx 2.65$). In a simplified assumption, the critical shear stress controls bedload entrainment, or even debris-flow formation at very steep slopes [Prancevic et al., 2014] and thus the bed composition and stability. Typically, bed stability is increased at steeper slopes and thus positive correlations of the critical shear stress with bed slope were identified in several studies [e.g., Bunte et al., 2013; Camenen, 2012; Ferguson, 2012; Lamb et al., 2008; Prancevic et al., 2014; Recking, 2009; Shvidchenko et al., 2001]. These positive correlations of $\tau^*_c$ against bed slope are mainly explained by increased bed stability at steep slopes due to interlocking of bed particles [Church et al., 1998], variable friction angles, flow aeration (lower water density) and increased turbulent conditions [Lamb et al., 2008]. However, in addition to the ratio between the available shear stress and the critical shear stress, there are other important factors in steep and narrow streams with a pronounced step-pool morphology: for example, the jamming of large particles, and thus the stability of steps, depends also on the sediment concentration in the flow and on the ratio of stream width to grain diameter [Church and Zimmermann, 2007].
Stream bed roughness and flow velocity

Here we study a steep, glacier-fed mountain stream with almost flume-like boundary conditions. Along the channel, the bed gradient systematically increases from 2.8% to 41%, while channel width and flow discharge remain approximately constant in the downstream direction during the summer period, which has the largest glacier meltwater runoff. We compiled a detailed dataset of bed characteristics (e.g., grain size distributions, bed topography) and flow velocities measured at different flow discharges. Based on these field data, we evaluated how well flow resistance equations and hydraulic geometry relations predict flow velocity at very steep (and to date almost unmeasured) bed slopes. We also compared different roughness measures for scaling flow depth and hydraulic geometry relations, using \( D_{84} \) and terrestrial laser scanning (TLS) point cloud statistics such as the standard deviation, the semi-variance or inter-percentile ranges of detrended bed elevations. Finally, we tested how bed roughness should adjust to bed slope and driving hydraulic forces, by combining the variable power flow resistance equation with recently derived empirical critical shear stress approaches, and comparing the results with measured bed characteristics.

2. Data and methods

2.1. Study site

The Riedbach is an alpine, glacier-fed stream located in the Matterhorn valley close to St. Niklaus (Figure 1). The Riedbach flows with a slightly meandering path over the glacier forefield with bed gradients ranging from about 2.8 to 6% before reaching a knickpoint and plunging into a very steep reach with gradients of up to 41% (Figures 2 and 3). In this stretch of the channel, the Kraftwerke Mattmark AG (KWM) operates a water intake for hydropower generation at an elevation of 1800 m a.s.l. At the water intake, located at the downstream end of the entire 1 km study reach, the Riedbach drains a watershed of 15.8 km\(^2\), of which 53% is glacier-covered. The catchment elevation range is 1800-4300 m a.s.l.

The runoff regime is dominated by snow and glacial melt during early and mid-summer. In this season, daily peak discharges range roughly between 2 m\(^3\)/s and 4 m\(^3\)/s (peak flows within that range are reached during about 50 days each year; Figure S1, supplementary material). Flow discharge rates are presumably higher following intense summer rainfalls, but the capacity of the water intake is about 4 m\(^3\)/s and any higher flows are unmeasured. In comparable alpine catchments in the Valais and southern Switzerland, the mean annual specific peak flow discharges are about 0.5 m\(^3\)km\(^{-2}\)s\(^{-1}\) with maximum values recorded ranging between 1-1.5 m\(^3\)km\(^{-2}\)s\(^{-1}\) [Weingartner et al., 2014]. From this range of specific peak flows and a catchment area of 16 km\(^2\) one can estimate that peak flows driven by intensive rainfall events may range from 8 to 24 m\(^3\)/s in the Riedbach.

Monitoring of sediment transport with geophone sensors indicates that transport primarily occurs in summer, when flow discharge rates are high due to glacial meltwater. During the 6-year period that these geophone sensors have been in operation (2009-2014), no debris flows were recorded in the main channel, suggesting that sediment transport in the Riedbach is predominantly characterized by fluvial erosion and transport. There is clear visual evidence of high sediment availability in the recent glacier retreat area, about 1 km above the beginning of study reach (see Figure S2, supplementary material). However, within the 1 km study reach, sediment availability and sediment supply appear to be low.
Chapter II

Figure 1: Riedbach catchment (Switzerland, Ct. Valais) and studied 1km stream reach. (© 2014 Google, Image Landsat, Image © 2014 DigitalGlobe). The glacier forefield is characterized by bed gradients $S$ ranging roughly from 3 to 12%; the steep reach is characterized by bed gradients ranging from 30 to 40%.

Figure 2: Study reaches with a) $S=2.8\%$ during low-moderate flow ($Q \approx 0.5 \text{ m}^3\text{s}^{-1}$), b) $S=12\%$ during moderate flow ($Q \approx 1.5 \text{ m}^3\text{s}^{-1}$), c) $S=30\%$ during low flow ($Q \approx 0.2 \text{ m}^3\text{s}^{-1}$), and d) $S=41\%$ during high flow ($Q \approx 3 \text{ m}^3\text{s}^{-1}$).
The glacier forefield in the study reach is vegetated and the banks are characterized by low gradients (Figure 2a and b), so significant sediment supply from bank erosion is unlikely. In the steep reach, the banks are mainly characterized by exposed bedrock (on the orographic left hand side of the channel) or dense vegetation (Figure 2d). There is only one tributary with regular debris-flow activity along the study reach; however its confluence is located immediately upstream of the water intake (on the orographic right hand side of the channel), and thus it does not influence the grain size distribution along the channel bed in the study reach.

For this study we defined channel reaches with more-or-less homogeneous bed gradients and bed morphologies, for which the bed characteristics (grain size distributions, bed topography) and the flow velocity and other hydraulic parameters were determined. At the upstream and downstream end of each reach, a fluorometer was installed for dye tracer studies. Study reaches and fluorometer locations were numbered as, for example, R#01 and Fl#01 respectively, as described in Table 1.

### 2.2. Characteristic grain sizes and roughness density

Grain size distributions were determined using the line-by-number (LBN) technique [Fehr, 1987], consisting of 290-1350 (median 380) counts of individual particles within each sampled reach. The characteristic grain sizes $D_{30}$, $D_{50}$, and $D_{84}$ coarsen significantly along the stream with increasing bed gradients from about 0.05/0.09/0.21 m to 0.2/0.43/1.18 m (Table 1). In addition, a grid-by-number pebble count (PC) was taken on the glacier forefield for R#01 [Bunte et al., 2013]. Full grain size distributions are given in the supplementary material (Figure S3).

Roughness density was characterized using the boulder concentration $\lambda=n_b\pi D_b^2/(4A_b)$, where $n_b$ is the number of boulders, $D_b$ is their mean diameter, and $A_b$ is the sampled area. As a simple assumption a fixed critical diameter of 0.5 m was used to define an immobile boulder, instead of using a critical diameter that depended on flow stage and bed gradient [cf. Nitsche et al., 2012].
### Table 1: Reach characteristics

<table>
<thead>
<tr>
<th>Reach Location</th>
<th>Reach ID</th>
<th>Slope [m/m]</th>
<th>Length [m]</th>
<th>Grain size distribution (LBN/PC)</th>
<th>Boulder density &gt;0.5m</th>
<th>Patches</th>
<th>TLS point cloud statistics</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>D&lt;sub&gt;10&lt;/sub&gt; [m]</td>
<td>D&lt;sub&gt;50&lt;/sub&gt; [m]</td>
<td>D&lt;sub&gt;84&lt;/sub&gt; [m]</td>
<td>D&lt;sub&gt;90&lt;/sub&gt; [m]</td>
</tr>
<tr>
<td>Fl#01-Fl#02</td>
<td>R#01</td>
<td>0.028</td>
<td>52</td>
<td>0.05/0.02</td>
<td>0.09/0.06</td>
<td>0.21/0.16</td>
<td>0.24</td>
</tr>
<tr>
<td>Fl#02-Fl#03</td>
<td>R#02</td>
<td>0.04</td>
<td>185</td>
<td>0.05</td>
<td>0.10</td>
<td>0.27</td>
<td>0.39</td>
</tr>
<tr>
<td>Fl#03-Fl#04</td>
<td>R#03</td>
<td>0.06</td>
<td>68</td>
<td>0.05</td>
<td>0.09</td>
<td>0.23</td>
<td>0.29</td>
</tr>
<tr>
<td>Fl#03-Fl#05</td>
<td>R#04</td>
<td>0.10</td>
<td>215</td>
<td>0.08&lt;sup&gt;(3)&lt;/sup&gt;</td>
<td>0.15&lt;sup&gt;(3)&lt;/sup&gt;</td>
<td>0.40&lt;sup&gt;(3)&lt;/sup&gt;</td>
<td>0.53&lt;sup&gt;(3)&lt;/sup&gt;</td>
</tr>
<tr>
<td>Fl#04-Fl#05</td>
<td>R#05</td>
<td>0.12</td>
<td>147</td>
<td>0.09</td>
<td>0.18</td>
<td>0.48</td>
<td>0.65</td>
</tr>
<tr>
<td>Fl#05-Fl#06</td>
<td>R#06</td>
<td>0.32</td>
<td>82</td>
<td>0.16</td>
<td>0.34</td>
<td>0.93</td>
<td>1.66</td>
</tr>
<tr>
<td>Fl#06-Fl#07</td>
<td>R#07</td>
<td>0.40</td>
<td>50</td>
<td>0.20</td>
<td>0.43</td>
<td>1.18</td>
<td>1.39</td>
</tr>
<tr>
<td>Fl#08-Fl#09</td>
<td>R#08</td>
<td>0.41</td>
<td>117</td>
<td>0.16</td>
<td>0.36</td>
<td>0.93</td>
<td>1.14</td>
</tr>
<tr>
<td>Fl#08-Fl#10</td>
<td>R#09</td>
<td>0.40</td>
<td>228</td>
<td>0.16&lt;sup&gt;(4)&lt;/sup&gt;</td>
<td>0.37&lt;sup&gt;(4)&lt;/sup&gt;</td>
<td>0.95&lt;sup&gt;(4)&lt;/sup&gt;</td>
<td>1.16&lt;sup&gt;(4)&lt;/sup&gt;</td>
</tr>
<tr>
<td>Fl#09-Fl#10</td>
<td>R#10</td>
<td>0.38</td>
<td>111</td>
<td>0.16</td>
<td>0.39</td>
<td>0.99</td>
<td>1.17</td>
</tr>
</tbody>
</table>

1) A<sub>b</sub> and A<sub>TLS</sub> = sampled area; λ = boulder concentration; IPR<sub>90</sub> = inter-percentile range based in the 95% percentile minus the 5% percentile; c' = semi-variogram sill level; STD<sub>z</sub> = standard deviation of the TLS point clouds.

2) LBN = Line-by-number sampling, PC = pebble count (only available for R#01).

3) Average from R#03 and R#05, weighted by reach length.

4) Average from R#08 and R#10, weighted by reach length.

5) Estimated from the measurement vs. bed slope relation: λ = 1.44*S<sup>1.12</sup> with r<sup>2</sup>=0.86.

6) Estimated from the measurement vs. bed slope relation: IPR<sub>90</sub> = 1.93*S<sup>0.74</sup> with r<sup>2</sup>=0.96 (See Figure 5a).

7) Estimated from the measurement vs. bed slope relation: c' = 0.76*S<sup>0.74</sup> with r<sup>2</sup>=0.96 (See Figure 5a).

8) Estimated from the measurement vs. bed slope relation: STD<sub>z</sub> = 0.59*S<sup>0.74</sup> with r<sup>2</sup>=0.66 (See Figure 5a).
2.3. Bed topography

The streambed surface was surveyed with a terrestrial laser scanner (Leica, C10, Leica Geosystems AG, Heerbrugg, CH) during autumn low-flow conditions. We scanned the reaches representative for (i) the glacier forefield with a 2.8% bed gradient (corresponding to the 50 m long Reach R#01), (ii) the transitional reaches with bed gradients from 6% to 42% (350 m long, Reaches R#03 to R#07), and (iii) the steep, downstream end of the study reach with a bed gradient of 38% (100 m long, Reach R#10) (see thick black lines in Figure 3). In total, we scanned from about 30 different positions resulting in an average point cloud density of about 5 points/cm$^2$, with a mean absolute registration error of 3 mm. Due to the very rough bed topography in the steep reach, shadow effects could not be avoided (Figure 4).

Cross-sectional profiles were derived from the TLS point cloud at 0.2 m intervals along the longitudinal profile for reaches R#01, R#03-R#07 and R#10. For reaches R#02 and R#08, cross-sectional profiles were recorded using a total station and a laser distance meter, respectively. From the complete point cloud, individual patches representing the local bed topography were exported for further statistical analysis (Table 1). The analyzed patches were defined according to the following criteria: no banks, as little missing data due to shading and water cover as possible, a representative area as large as possible, no large-scale structures (e.g. concave or convex forms, coves or notches), and a minimum width twice the largest grain size within each reach.

Figure 4: Point cloud patch examples within a) reach R#01, 2.8% bed slope, b) reach R#03, 6% bed slope, and c) reach R#07, 40% bed slope. Black areas represent shadows without data.
Chapter II

The analyzed patches were typically rectangular with two of their sides parallel to the flow direction. The noise within the point cloud was reduced and outliers were removed using Geomagic Studio 2014 (3D Systems, Rock Hill, SC, USA, 2014). The point cloud density was equally spaced and limited to not exceed 4 points/cm². Finally, the patches were detrended for local slope.

As a measure of bed roughness for each analyzed patch, we calculated the 90% inter-percentile range (IPR₉₀) defined by the 95th percentile minus the 5th percentile of all detrended elevation values. Furthermore, we calculated the standard deviation \( \text{STD}_z \) (see also Brasington et al. [2012]) and the semi-variogram [Hodge et al., 2009; Robert, 1988; 1991] using the de-trended elevations. For each patch within a reach with a characteristic bed gradient (R#01-R#10), the surface statistics (IPR₉₀, the semi-variogram sill values, and the standard deviation \( \text{STD}_z \)) of the TLS point cloud were averaged (median), and the variability between patches was characterized by the inter-patch standard deviation (note, this standard deviation is different from \( \text{STD}_z \); Table 1). In addition, the point cloud was interpolated onto a regular grid and the roughness parameters (the inter-percentile range, the standard deviation and the semi-variogram) were re-calculated based on the gridded data. This was done to further quantify the uncertainties in the roughness measures, whether derived from the point cloud or gridded data.

2.3.1 Semi-variance

The semi-variance is a commonly used statistic to characterize spatial correlations of vertical distances at increasing lags, and it is defined as half the variance at lag \( h \) (Equation 5).

\[
\gamma(h) = \frac{1}{2N(h)} \sum_{i,j \in N(h)} (z_i - z_j)^2
\]  

(Equation 5)

Here, \( N \) is the number of elements (within lag \( h \)) and \( z \) is the vertical elevation. The semi-variance of a gravel bed river with irregular structure is typically characterized by an increasing semi-variance with increasing lag distance until the semi-variance reaches a more-or-less constant level (the “sill”) at some lag distance (the “range”). To determine the range and sill values we fitted a simple spherical model (Equation 6; see Figure S4 supplementary material) to the isotropic semi-variogram:

\[
\gamma(h) = \begin{cases} 
    c_n + c_p \left( \frac{3h}{2r} - \frac{1}{2} \left( \frac{h}{r} \right)^3 \right) & \text{if } 0 \leq h \leq r \\
    c_n + c_p & \text{if } h > r 
\end{cases}
\]  

(Equation 6)

Where \( r \) is the range, \( c_p \) is the partial sill value and \( c_n \) is the nugget (defined here as the semi-variance at the smallest lag-distance). The total sill value is defined in this study as \( c = c_p + c_n \). For computational reasons, the semi-variogram was usually calculated on 150000 points. In preliminary tests it could be shown that including additional points did not increase the quality of the semi-variogram. The minimum lag distance was 0.01 m and the maximum lag was defined as half the width of the analyzed patch. Because the sill value \( c \) is related to the semi-variance, and variance is the square of the standard deviation, we multiplied \( c \) by two and took the square root \( c' = \sqrt{2c} \) to compare the sill with the other roughness measures, such as the characteristic grain sizes, inter-percentile ranges and standard deviation.
2.4. Flow velocity

During summer and autumn 2013, we conducted 46 dye tracer experiments at the Riedbach. Ten in-situ fluorometers (GGUN-FL30, Albillia SA; Schneeg [2003]) were installed along the study reach from the glacier forefield to the water intake. The positions were selected to define reaches with relatively homogeneous bed gradients and bed morphologies between pairs of fluorometers (Figure 3, Table 1). The fluorometers were installed as far as possible (typically around 1 m) from the banks. It was not possible to install fluorometers in the middle of the stream during high flows. Dye tracer was either injected on the glacier forefield at injection location Inj#01, 92 m upstream of the first fluorometer Fl#01, or near the transition to the steep reaches, at injection location Inj#02, 43 m upstream of Fl#05. The tracer was injected by splashing a bucket of diluted dye as evenly as possible across the width of the channel.

The raw fluorometer data (mV) were calibrated against reference dye concentrations (ppb), and measured background values were leveled to zero before each injection. The beginning of each individual break through curve (BTC) was determined visually. The end of the BTC was defined as the point where 99% of the tracer had passed the fluorometer. Outliers within the BTCs were removed using a robust spline filter. For each BTC the concentration-weighted harmonic mean was determined to derive the mean travel time and flow velocities [Waldon, 2004]. The reach-averaged flow velocity was determined by scaling the distance by the (harmonic) mean travel times between two fluorometers. To increase the database for further analysis, not only reaches between two subsequent fluorometers were considered but also reaches between all pairs of fluorometers (Table 1) without significant changes in bed slope and bed topography (Table S1 in the supplement includes the flow velocity data).

2.5. Flow discharge

Flow discharge rates $Q_{WI}$ were gauged in 10-minute sampling intervals in the settling basin of the water intake at the downstream end of the steep study reach. We assume that flow discharge rates are approximately constant along the 1 km study reach because of the glacial runoff regime and because no considerable rainfall events occurred during the field measurements. Also, it is assumed that there was no considerable exfiltration from ground water as estimated by integrating the area under the dye tracer BTC’s [Leibundgut et al., 2009] (a detailed explanation for this assumption can be found in Section S2, supplementary material). We used the same flow discharge rates for the 1 km long study reach as measured at the water intake, corrected for unmeasured residual and overtopping water (see Section S2, supplementary material). Throughout this study, flow discharge rates $Q$ refer to corrected values of $Q_{WI}$ measured at the water intake.

2.6. Hydraulic parameters

Hydraulic parameters including flow area ($A$), flow width ($w$), wetted perimeter ($w_p$), flow depth ($d_h=A/w$) and hydraulic radius ($d=A/w_p$) were back-calculated from the reach-averaged flow velocity ($v$) for each cross-section using the continuity equation ($A=Q/v$). Note, flow depth in all calculations and figures is approximated with the hydraulic radius. The hydraulic parameters were then averaged over the available cross-sections within each reach. Averaging the hydraulic parameters rather than the available cross-sections itself (as done by Nitsche et
was preferred due to the rough bed topography, including large boulders, which made it difficult to derive averaged cross-sections. Note that hydraulic parameters were back-calculated as reach averages and thus neglect potential water-air mixtures.

### 2.7. Prediction of $D_{84}$

To develop an equation to predict the increase of bed roughness, defined here by the $D_{84}$, for increasing bed gradients, we start with the idea of a governing or dominant flow discharge per unit width. We assume that the most relevant discharges for channel bed adjustments range between the frequently occurring high flows of about 3 m$^3$/s total discharge (or 0.4 m$^3$/s-1m$^{-1}$ unit discharge) due to glacier melt in mid-summer, and the individual extreme rainfall-driven peak flows which were estimated to reach 24 m$^3$/s, or 2.0 m$^3$/s-1m$^{-1}$ (Section 2.1; and mean hydraulic geometry relation for width as function of discharge in Table S2, supplementary material). We then used a flow resistance equation [Ferguson, 2007] (Equation A1) combined with critical shear stress approaches to solve for a critical grain diameter $D_{84}$. The detailed derivation and the resulting equation to estimate $D_{84}$ as a function of bed slope can be found in Appendix A. The critical shear stress approaches considered are derived from Lamb et al. [2008] (Equation 7) and Camenen [2012] (Equation 8), based on $\tau^*_{c0}$=0.05 and an angle of repose $\phi_s$ of 50°.

\[
\tau^*_c = 0.15S^{0.25}
\]

\[
\tau^*_c = \tau^*_{c0} \frac{\sin(\phi_s - \arctan S)}{\sin(\phi_s)} \left(0.5 + 6S^{0.75}\right)
\]

Because the critical shear stress approaches refer to a $D_{50}$, the equations were modified assuming a factor of $w = D_{84}/D_{50} =2.6$ (see Appendix, Equation A5), corresponding to the average $D_{84}/D_{50}$ ratio measured in the Riedbach. Using a value of $w = 2.6$ assumes that no hiding and exposure effects would ease particle entrainment for the $D_{84}$ size class compared to the $D_{50}$ size class. Alternatively, to account for potential hiding effects [Parker, 2008], the hiding function of Wilcock and Crowe [2003] was used, with $b = (0.67/1+\exp(1.5-2.6)) = 0.503$, giving $\tau^*_{c84}/\tau^*_{c50} = (D_{84}/D_{50})^{b-1} = 0.622$ for $D_{84}/D_{50} = 2.6$, resulting in a reduced factor $w' = 2.6*0.622$.

### 3. Results

#### 3.1. The channel bed morphology

With increasing bed gradients, a dramatic increase in the bed's grain size diameter and topographic irregularity can be observed. All measures of bed roughness show strong positive correlations with bed slope, with power law slopes around 0.75 (Figure 5a). Whereas the inter-percentile range $IPR_{90}$ is similar to $D_{84}$, the sill value $c'$ and the $STD_e$ are similar to $D_{50}$. The characteristic grain sizes $D_{50}$ and $D_{84}$ from reaches with shallow bed gradients (<6% bed slope; data points indicated by blue circles in Figure 5) deviate from the trend for bed gradients steeper than 6%, so the power laws trends were estimated for the steep reaches (>6%) only. Also, the characteristic grain sizes in reach R#01 derived from pebble counts by Bunte et al. [2013] are smaller than the characteristic grain sizes derived from line-by-number samples (Figure 5) (for more details on these observations see section 4.1).
Deviations in the roughness measures, whether derived directly from the point cloud directly or from interpolated, gridded data, are less than 3\% for the STD$_z$, 3\% for the IPR$_{90}$ and 17\% for the sill value $c'$ (Figure S5).

Figure 5: a) Roughness height related to channel bed slope; b), c) and d) selected roughness heights plotted against each other. $D_{64}$ and $D_{50}$ (squares and triangles, respectively) are based on line-by-number measurements. Red squares and triangles refer to the characteristic grain sizes derived from pebble counts (only available for R01). Blue large circles represent the strongest deviations between the point cloud statistics and the characteristic grain sizes, and were removed from the power-law fit. Surface statistics sill ($c'$, circles), standard deviation (STD$_z$, diamonds) and inter-percentile range (IPR$_{90}$, crosses) were calculated on de-trended elevations of individual point cloud patches. Plotted points are medians of 4-6 patches for each characteristic bed slope class (3\%, 6\%, 12\%, 30\%, 40\% and 41\%; see Table 1). Interpolated values given in Table 1 are neglected here. Error bars refer to the standard deviation. Note: bed slopes for $c'$ and IPR$_{90}$ points were shifted slightly to avoid overlaps between error bars.
3.2. Flow velocity, depth and width related to bed slope

Measured flow velocities of all reaches obey clear power-law relationships with measured flow discharge ($r^2 \geq 0.97$; Table 2). Flow velocity is slightly higher on the glacier forefield (at bed slopes shallower than 6%) than it is in the steeper reaches (≥6%), for which flow velocity is nearly constant with increasing gradient, as shown in Figure 6a for 0.5, 1.0, 2.0, and 3.0 m$^3$/s total discharge. Flow depth and width are back-calculated from reach-averaged velocity and the cross-sectional profiles, and show no clear trends with increasing bed slope (Figure 6, b and c). The smaller flow width in reach R#01 compared to R#02 and R#03 on the glacier forefield corresponds to field observations of a relatively narrow channel at this reach. Unit discharge rates $q=Q/w$ (m$^3$/s$^{-1}$) were determined for each reach and each dye tracer experiment based on the flow width relation $w=aQ^b$ (Table S2, supplementary material).

Flow velocity for unit discharges of 0.1, 0.2, 0.3 and 0.4 m$^3$/s$^{-1}$ (corresponding roughly to 0.5, 1.0, 2.0, and 3.0 m$^3$/s total discharge) is generally highest on the glacier forefield (at bed slopes of S=3-12%) and somewhat smaller for the steeper reaches with S≥12% (Figure 6d).

![Figure 6](image_url)

Figure 6: a) Reach-averaged flow velocity, b) reach-averaged flow depth and c) reach-averaged flow width related to channel bed slope at different flow discharge rates $Q$. Blue shadow shows uncertainties derived from the fit of the hydraulic geometry relations (standard error $SE$, Table 2 and S2) for the highest discharge; uncertainties for the other discharges would be similar. d) Flow velocity for unit discharge rates $q=Q/w$ related to bed slope (in average $q=0.1, 0.2, 0.3, 0.4$ m$^3$/s$^{-1}$ correspond to roughly $Q=0.5, 1, 2$ and 3 m$^3$/s).
### Table 2: Power law relations for velocity vs. discharge; Darcy-Weisbach and dimensionless hydraulic geometry relations (Figures 7 and S6).

<table>
<thead>
<tr>
<th></th>
<th>$v=kQ^n$</th>
<th>Darcy-Weisbach: $(8/\pi a^3)=a(dR)^b$</th>
<th>Diml. Hydr. Geom.: $v^{<strong>}=aq^{</strong>}$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$k$</td>
<td>$m$</td>
<td>$r^2$</td>
</tr>
<tr>
<td>R#01</td>
<td>0.95</td>
<td>0.58</td>
<td>0.97</td>
</tr>
<tr>
<td>R#02</td>
<td>0.86</td>
<td>0.54</td>
<td>1</td>
</tr>
<tr>
<td>R#03</td>
<td>0.67</td>
<td>0.51</td>
<td>0.99</td>
</tr>
<tr>
<td>R#04</td>
<td>0.69</td>
<td>0.54</td>
<td>1</td>
</tr>
<tr>
<td>R#05</td>
<td>0.73</td>
<td>0.55</td>
<td>1</td>
</tr>
<tr>
<td>R#06</td>
<td>0.7</td>
<td>0.59</td>
<td>0.99</td>
</tr>
<tr>
<td>R#07</td>
<td>0.7</td>
<td>0.61</td>
<td>0.97</td>
</tr>
<tr>
<td>R#08</td>
<td>0.88</td>
<td>0.54</td>
<td>0.98</td>
</tr>
<tr>
<td>R#09</td>
<td>0.86</td>
<td>0.49</td>
<td>0.99</td>
</tr>
<tr>
<td>R#10</td>
<td>0.85</td>
<td>0.44</td>
<td>0.98</td>
</tr>
<tr>
<td>Mean</td>
<td>0.8</td>
<td>0.54</td>
<td>0.99</td>
</tr>
<tr>
<td>STD</td>
<td>0.09</td>
<td>0.04</td>
<td>0.01</td>
</tr>
</tbody>
</table>
Uncertainties in flow velocity measurements due to incomplete dye tracer mixing were quantified by installing two fluorometers across a cross-section. Uncertainties are maximum 13% during moderate flows in the first reaches downstream of each injection location (Inj#01 and Inj#02), and 1-5% during low flows in the second reaches downstream of each injection location. Uncertainties during high flows could not be quantified, but are assumed to be higher. In the analysis presented here, uncertainties in the flow velocity estimates are assumed to not substantially affect the outcome of the results.

3.3. Relating flow velocity to bed characteristics

To explain the observed flow velocities and relate them to measured bed characteristics, two commonly used concepts were considered, the Darcy-Weisbach relation (Equation 1) and the dimensionless hydraulic geometry relations (Equation 2 and 3). The bed roughness height measures include \(D_{84}\) and the point cloud statistics, i.e. inter-percentile range \(IPR_{90}\), semi-variogram sill value \(c'\) and the standard deviation \(STD_z\). Because \(IPR_{90}\) shows a strong correlation to \(D_{84}\), and \(STD_z\) to \(c'\), respectively (Figure 5), results for \(IPR_{90}\) and \(STD_z\) are shown in the supplementary material (Figure S6), whereas results for \(D_{84}\) and \(c'\) are shown here (Figure 7).

3.3.1 Darcy-Weisbach coefficient

Back-calculated roughness coefficients \((8/f_{tot})^{0.5}\) based on measured flow velocity vary systematically with relative flow depth \(d/R\) (Figure 7a, b). For all reaches, positive correlations could be identified with power law exponents ranging from roughly 1 to 2 and correlation coefficients \(r^2\) from 0.63 to 0.98 (Table 2). Power laws, shown by the thick black lines in Figures 7a, b, were fitted to the collapsed data of each roughness measure, taking the different number of data points per reach into account. The similarity collapse of the individual reach-wise relations is generally comparable for all of the considered roughness measures, i.e. the \(D_{84}\) and the sill value \(c'\) (Figure 7a, b) as well as the \(IPR_{90}\) and the standard deviation \(STD_z\) (Figure S6). All four roughness measures give comparable fits, as measured by their coefficients of determination \(r^2\) and the root mean square errors RMSE. However, a real collapse between the reaches is not possible for any of the roughness measures, because the power law exponents of the individual reaches are considerably steeper than the exponents fitted jointly to all the data points.

The Riedbach data plot within the upper range of the data compiled by Rickenmann and Recking [2011] at comparable relative flow depths (Figure 7a). The reaches with larger relative flow depths (glacier forefield, R#01 and R#02) follow the relations of the reaches with smaller relative flow depths, although none of these relative flow depths are large (much greater than 1). The Ferguson [2007] equation (A1) was fitted to the Riedbach data (red dashed line, Figure 7a) by optimizing the parameters \(a_1\) and \(a_2\) based on Monte-Carlo simulations (Figure S7). The parameter \(a_2\) is most relevant for low relative flow depths, defined here as \(d/R<1\); the value of \(a_2=3.9\) was obtained from Monte Carlo simulations (see Fig. S7).
Figure 7: Relationship between $(8/f_{tot})^{0.5}$ and relative flow depth scaled by a) the $D_{84}$ derived from line-by-number samplings and b) the sill level $c'$ of the de-trended laser scan point clouds. Black thick lines show a power law fit to the data of all reaches. Relationships for individual reaches (thin colored lines) are steeper; fit details are given in Table 2. c) Dimensionless velocity related to dimensionless discharge scaled by the $D_{84}$ and d) the sill level $c'$. Gray dots in a) show the data of Rickenmann and Recking [2011] (RR2011). The equation of Rickenmann and Recking [2011] is shown by the black dot-dashed line. In addition Equation A1 (Ferguson [2007], Ferg2007) is shown both with the standard parameter set of $a_1$ and $a_2$ (black dashed line) and the optimized parameter set (red dashed line, more details see text). The optimized equation A1 was used for the calculations presented in Fig. 9. In d) the equation given in Yochum et al. [2012] is shown.
Figure 8: Measured flow velocity compared to predictions based on scaling flow depth $d$ with a roughness measure $R$ (Darcy-Weisbach, left side of plot), and compared to predictions based on scaling flow velocity and discharge (dimensionless hydraulic geometry, right side of plot). Equation A1 refers to the VPE-Equation of Ferguson [2007]. Equation 21b and 22 are given in Rickenmann and Recking [2011] (RR2011). Optimized dimensionless hydraulic geometry equations for the Riedbach data are based on the fitted overall relations (black thick lines) given in Figure 7 ($R=D_{84}$ and $R=c'$) and Figure S6 ($R=IPR_{90}$ and $R=STD_z$), respectively.

Because almost no data are available for $d/R>1$ (for which the parameter $a_1$ becomes more relevant), $a_1$ was not sensitive to the Monte Carlo simulations (Figure S7) and was therefore manually set to $a_1=9.5$. Choosing $a_1=9.5$ results in an equation for $d/R>1$ that basically follows the upper part of the Rickenmann and Recking [2011] data, similarly as the Riedbach data do for $d/R<1$.

The Rickenmann and Recking [2011] equation, the original Ferguson [2007] equation (Equation A1, $a_1=6.5$ and $a_2=2.5$), the fitted Ferguson [2007] equation (Equation A1, $a_1=9.5$ and $a_2=3.9$), as well as empirical relations fitted to the collapsed data (thick black lines in Figures 7 and S6) were used for flow velocity prediction (Figure 8). As expected from Figure 7, the Rickenmann and Recking [2011] equation and the original Ferguson [2007] equation underestimate flow velocity slightly, whereas the fitted Ferguson [2007] equation and the fitted power law equations (thick black lines in Figures 7 and S6) generally perform better (Figure 8). The choice of the roughness measures does not affect the performance of the equation.

3.3.2 Dimensionless hydraulic geometry

Flow velocity and unit discharge of selected reaches were non-dimensionalized using the different roughness heights $R$ (Equation 2 and 3). The hydraulic geometry relations of the
different reaches are all very well defined with power law exponents ranging between 0.6-0.7 and with $r^2$ values larger than 0.98 (Figure 7c, d; Table 2).

The collapse of the individual reach-wise relations is generally better defined for the hydraulic geometry relations (Figure 7c, d) than for the Darcy-Weisbach relations (cf. RMSE values in Figure 7a, b). It should be noted that this is partly an artificial result because the data are now spread over almost two orders of magnitude as compared to only one order of magnitude in plots of $(8/f_{rot})^{0.5}$ vs. $d/R$. The performance of the collapse again is comparable for all considered roughness measures (Figure 7c, d; Figure S6 e, d).

Similarly as observed for the Darcy-Weisbach relations (Figure 7a), most of the Riedbach data plot in the upper part of the data compiled by Rickenmann and Recking [2011] (Figure 7c). The average exponent fits very well to the exponent of 0.705 given by Rickenmann and Recking [2011] for large-scale roughness. The fitted equation of the dimensionless hydraulic geometry relations based on the sill value $c'$ is very close to the relation presented in Yochum et al. [2012] (Figure 7b). Furthermore, the Riedbach data of the steep reaches (R#06-R#10) with small relative flow depths (which are under-represented in the Rickenmann and Recking [2011] data), follow the trends of the reaches with larger relative flow depths. However, the physical interpretation of these values for very low relative flow depths ($d/R<0.5$) might be problematic if part of the channel width is occupied by larger, flow protruding grains [cf. Rickenmann and Recking, 2011]. The predictive accuracy for the dimensionless hydraulic geometry relations is slightly improved as compared to the Darcy-Weisbach approach. Again, the choice of the roughness measures does not affect the performance of the equations (Figure 8).

3.4. Adjustment of bed roughness to channel slope and hydraulic conditions

The critical shear stress approaches of Camenen [2007] and Lamb [2008] (Equations 7 and 8) were combined with the flow resistance equation of Ferguson [2007] adjusted to the Riedbach flow velocity measurements (Equation A1 with $a_1=9.5$ and $a_2=3.9$) to predict the characteristic grain size $D_{84}$ as a function of bed slope (Equation A7, Figure 9). Using Equation A1 is justified because it has a good overall performance in describing the measured flow velocities for various channel slopes. We assumed a dominant unit discharge range of $q=0.4-2$ m$^3$m$^{-1}$s$^{-1}$ (corresponding on average for the entire study reach to 3-24 m$^3$m$^{-1}$ total discharge), considered to be responsible for the formation of the bed roughness, and we also considered potential hiding effects. Both the Camenen [2007] and Lamb [2008] equations correctly predict the slope of the relationship between $D_{84}$ and bed gradients observed in the Riedbach (Figure 9). However, for any particular bed gradient, the predicted values of $D_{84}$ vary by almost an order of magnitude, depending on whether the Camenen [2007] and Lamb [2008] equation is used and depending on whether hiding effects are considered or not (Figure 9); varying the dominant discharge between 0.4 and 2 m$^3$m$^{-1}$s$^{-1}$ introduces a further factor of $\sim$3 variation in predicted $D_{84}$. Assuming a dominant discharge of 0.4 m$^3$m$^{-1}$s$^{-1}$ and neglecting potential hiding effects provides very good agreement of the Camenen [2012] approach with field observations over the entire bed slope range, whereas the Lamb et al. [2008] equation slightly underestimates observed $D_{84}$ values at lower bed gradients (Figure 9a). Accounting for potential hiding effects results in a strong underestimation of the predicted $D_{84}$ for both approaches (Figure 9a). By contrast, assuming very high flows (5-6 times bankfull) with a
unit discharge of 2 m$^3$/s$^{-1}$ and accounting for potential hiding effects provides very good estimates of $D_{84}$, whereas neglecting hiding effects leads to a strong overestimation of predicted $D_{84}$ values (Figure 9b).

![Graph](image)

Figure 9: Predicted $D_{84}$ (colored lines) compared to measured $D_{84}$ values (squares). Predictions are based on the critical shear stress approaches of Camenen [2012] (dark blue dashed and solid lines) and Lamb et al. [2008] (light blue dashed and solid lines) combined with the flow resistance equation of Ferguson [2007] ($a_1=9.5$, $a_2=3.9$; Equation A7). Solid lines neglect potential hiding effects, whereas dashed lines consider hiding effects based on the Wilcock and Crowe [2003] hiding function. a) A bed controlling unit discharge of 0.4 m$^3$/s$^{-1}$ (about 3 m$^3$/s total discharge on average, occurring frequently during summer high flows) is assumed b) A bed controlling unit discharge of 2 m$^3$/s$^{-1}$ (about 24 m$^3$/s total discharge, representing rare intense storm events) is assumed.
4. Discussion

4.1. Bed roughness characterization

The increase of the Riedbach bed slope is associated with a considerable increase in bed roughness, as shown for the different roughness measures \(D_{50}, D_{84}, IPR_{90}, c'\) and \(STD_z\), with power law exponents ranging roughly around 0.75 (Figure 5). The line-by-number (LBN) samples at shallow bed slopes (\(S<6\%\); blue circles, Figure 5) deviate from the power-law trends of the steeper reaches, which might be explained by sampling uncertainties arising from the underestimation of fine (and therefore uncountable) gravels. This potential LBN sampling bias is consistent with the smaller values for \(D_{50}\) and \(D_{84}\) derived from a pebble count in reach R#01 [Bunte et al., 2013]. Despite the uncertainties in the \(D_{50}\) and \(D_{84}\) values for \(S<6\%\), the regression lines for \(S>6\%\) are parallel to the relations of the other, statistically better-verified roughness measures (Figure 5a). This suggests that even if characteristic grain sizes are derived from (potentially biased) LBN surveys, they are adequate for determining the characteristic roughness height.

The standard deviations of the point cloud statistics (\(IPR_{90}, STD_z, c'\)) in Table 1 range between 10 and 60\% of the median values for each reach. These variations between the individual patches may represent the spatial variability of streambed roughness, or the difficulty in identifying representative patches, especially for the 12\% reach where the maximum variability of 60\% arises. Within-patch uncertainties in the surface statistics, whether derived from the point cloud directly or from gridded data, are 3-17\% (Figure S5), and thus are relatively small compared to the between-patch variations (see error bars in Figure 5). Therefore, the challenging task of meshing the point clouds of complex bed topographies such as in the steep part of the Riedbach does not convey clear advantages, in terms of better-constrained roughness estimates, compared to roughness estimates derived directly from the point clouds. In rough beds with shadow effects, the standard interpolation algorithms add points in the shaded areas on the level of the lowest surrounding points, and therefore underestimate the depth of the pockets between roughness elements. Therefore, all roughness measures, whether derived from the point cloud or meshed data, can be assumed to underestimate true roughness, especially at steep slopes, where shadow effects are larger. Recent developments in airborne laser scanning or drone-based photogrammetry, as well as developments of algorithms to identify the geometry of individual grains from the existing point clouds, are not considered in this study but would provide new opportunities for reducing shadow problems in mapping rough surfaces.

4.2. Flow resistance and velocity

The results from the flow velocity measurements raise the question how flow energy is dissipated so effectively that flow velocity is constant (or even slows down between reaches R#01 and R#03) with increasing bed gradients (Figure 6). Intuitively, one may assume that the reason is an increase in bed roughness and thus flow resistance (grain, form and spill resistance). The decreasing velocities from the glacier forefield to the steep reach might be explained by increasing spill resistance. Reach-averaged Froude numbers \(Fr>1\) were rare, occurring only in reaches with \(S<6\%\) and only during the highest flows (Figure S8). In the steep reaches (\(S>10\%\)), however, low relative flow depths \(d/R\) (cf. Figure 7a, b) imply that the
flow could be locally supercritical and thus could generate considerable spill resistance (although the reach-averaged Froude numbers were always subcritical in these reaches).

To compare the measured flow velocity and bed characteristics, we related the Darcy-Weisbach coefficient in the form of \((8/f_{tot})^{0.5}\) to the scaled flow depth (Figure 7a, b), and similarly related the scaled velocity \(v^{**}\) to the scaled unit discharge \(q^{**}\) (dimensionless hydraulic geometry; Figure 7c, d). Both \((8/f_{tot})^{0.5}\) and the hydraulic geometry relations of the different reaches can be approximately collapsed onto single curves. The hydraulic geometry relations (Figure 7c, d; Figure S6c, d) provide an improved collapse compared to the Darcy-Weisbach relations (Figure 7a, b; Figure S6a, b) as measured by the \(r^2\) and RMSE values of the power-law fit to the entire data. Also, the dimensionless hydraulic geometry relations provide better flow velocity predictions than the Darcy-Weisbach relations do (Figure 8). The observation that unit discharge provides a more robust measure than flow depth of the stress acting on the streambed has previously been reported [e.g., Recking, 2010; Rickenmann and Recking, 2011; Yochum et al., 2012]. Highly accurate flow velocity prediction based on the concept of dimensionless hydraulic geometry relations has also been reported previously [e.g., Comiti et al., 2009; Nitsche et al., 2012; Rickenmann and Recking, 2011; Yochum et al., 2012; Zimmermann, 2010]. However, the better collapse for the dimensionless hydraulic geometry relations is also partly an artificial result of stretching the data on both axes, because the common scaling factors (bed slope \(S\) and roughness length \(R\)) are more variable than the original dimensional quantities, discharge \(q\) and velocity \(v\). It should also be noted that the unit discharge depends on flow width \((Q/w)\), which in turn is calculated from measured flow velocity \((w=f(Q/v))\), where \(f\) depends on the shape of the channel cross-section. Thus the apparently better flow velocity predictions obtained by the dimensionless hydraulic geometry approach could be partly artifactual.

The collapse of the individual stream reach relations onto an overall equation is comparable for all considered roughness measures \((D_{94}, IPR_{90}, c', STD_z \text{ Figure 7 and S6})\). Thus, despite the various problems in determining a characteristic grain size (see literature review in introduction), the collection of high-resolution TLS data may not improve flow velocity predictions (Figure 8) enough to justify the cost and workload involved. Furthermore, if high-resolution TLS data are available, the computationally more complex calculation of the semi-variances does not appear to significantly improve the flow resistance analysis, compared to the more straightforward calculation of the standard deviation of detrended bed elevation.

However, no perfect similarity collapse of the flow resistance data was obtained using any of the roughness measures, which is more evident for the Darcy-Weisbach presentation (Figure 7). Fitting power laws with a fixed average exponent of 1.3 (see Table 2: mean \(b\) value in column “Darcy-Weisbach, \(c'\)” (Figure 7b)) results in different prefactors \(k\), which are positively correlated with bed slope (Figure 10a). These differences in flow resistance might be explained by variable densities of large-scale roughness elements in the Riedbach. With increasing boulder density (of up to 40% in the Riedbach), these roughness elements might begin to interfere with each other, resulting in skimming flows [see Lettau, 1969; Smith, 2014]. In the Riedbach, there is a positive correlation of the prefactor \(k\) with the roughness density (Figure 10a).
Stream bed roughness and flow velocity

Figure 10: Prefactors \( k \) derived from fitted power laws with a fixed average exponent to the reach-wise relations shown in Figure 7b, related to a) bed slope and b) roughness density. Circles refer to measured values; crosses refer to interpolated values (see Table 1). Red line was fitted to measured values only.

This relation stands in contrast to the findings of Nitsche et al. [2012] who reported negative trends for a very similar analysis, but one that compared flow resistance between streams, rather than among reaches of one particular stream. These results may not necessarily contradict each other, however, because the roughness densities reported by Nitsche et al. [2012] were less than 10%, except in the two steepest reaches. It seems reasonable that within that range, increasing roughness density should generally increase flow resistance.

4.3. Adjustment of bed roughness to flow conditions

As a simplifying assumption, the critical shear stress (Equation 4) can be considered as a parameter describing the interactions between the water flow, channel bed and bedload transport. Empirical critical shear stress approaches were combined with a flow resistance equation to predict the characteristic grain size \( D_{84} \) as a function of bed slope and the dominant flow conditions (Figure 9). The frequent summer high flows due to glacial melt (\( Q \geq 3 \text{ m}^3\text{s}^{-1} \)) and episodic high flows due to intensive rainfall events (\( Q \leq 24 \text{ m}^3\text{s}^{-1} \)) were considered to be responsible for the evolving bed roughness. Both the Lamb [2008] and Camenen [2012] approaches for estimating critical shear stress (Equations 7 and 8) reproduce the increase in \( D_{84} \) with increasing bed gradients (Figure 9), although the absolute values of the predicted \( D_{84} \) strongly depend on the choice of the dominant unit discharge and on the presence or absence of hiding effects. However, the estimates of the dominant discharge and the potential hiding effects both appear to be realistic.

The different predictions resulting from the Camenen [2012] and the Lamb [2008] approaches can be explained by the form of the original equations. While the Camenen [2012] equation approaches a constant critical shear stress \( \tau^*_{c} \) at very low bed gradients (\( \tau^*_{c}=0.05 \) is used here) the Lamb [2008] equation is based on a simple power law resulting in considerably lower \( \tau^*_{c} \) and therefore lower predicted \( D_{84} \) values, at low bed gradients (Figure 9).

The close agreement between predicted and measured \( D_{84} \) values indicates that the empirical critical shear stress approaches of Camenen [2012] and Lamb [2008] are valid also in the Riedbach. Our results also support Prancevic et al.’s [2014] physical explanation for
increasing critical shear stress with increasing bed slope, which was mainly based on flume experiments. The good performance of the critical shear stress relation of Camenen [2012] suggests that in the Riedbach, the feedback between bed roughness, flow velocity, and sediment transport is probably close to steady state conditions. Bedload transport is not the focus of this study, however annual sediment volume estimates were made both for the glacier forefield and for the downstream end of the steep study reach, and they support the assumption of a stable bed and a system in approximate steady state conditions [Schneider et al., 2013].

The stability and adjustment of steep, narrow streambeds like the Riedbach may be influenced by other factors in addition to the bed shear stresses considered here. For example, ratios of stream width to boulder diameter and sediment transport concentrations have been suggested to be important controls on the stability of step-pool channels [Church and Zimmermann, 2007]. However, we assume that these factors play a minor role in the Riedbach. Stream width is not predefined by the local topography (except for bedrock outcrops on the left bank in the steep reaches) and therefore is freely adjustable as well. Furthermore, the sediment transport concentration, which is assumed to absorb flow energy during transport, should be comparable in all of the reaches because the main sediment supply to our study reach is from the main channel upstream in the glacier retreat area.

5. Conclusions

We studied flow velocity, channel roughness, and bed stability at a steep mountain stream spanning a wide range of bed gradients. The study reach of the Riedbach forms an interesting setting with almost flume-like boundary conditions. The bed gradient increases by roughly one order of magnitude over only 1 km stream length, while the discharge and flow width remain approximately constant. Detailed field measurements of flow and bed characteristics led to the following conclusions:

(i) Bed roughness increased systematically as the ~0.75 power of bed gradient (Figure 5). Despite some uncertainties in the line-by-number grain size measurements for the reaches with smaller roughness heights (on the flat glacier forefield), the characteristic grain sizes derived from this easily applied method are in good agreement with the statistically better supported roughness measures derived from TLS point cloud data. Roughness heights estimated from the $D_{84}$ and from TLS point clouds provided comparable flow velocity predictions.

(ii) Flow velocity was faster on the glacier forefield ($S<6\%$) and slower in the steep reaches ($S=6-41\%$, Figure 6) with greater bed roughness (as measured by $D_{84}$, $IPR_{90}$, $c'$ and $STD_{z}$). Both the Darcy-Weisbach relation (scaling flow depth) and the closely related concept of dimensionless hydraulic geometry (scaling flow discharge) predicted the measured flow velocities for a wide range of bed gradients, including very steep slopes (Figures 7 and 8). However, somewhat better results were obtained when using the hydraulic geometry approach. These findings are in a close agreement with previous studies that focused on lower-gradient streams [e.g., Ferguson, 2007; Nitsche et al., 2012; Rickenmann and Recking, 2011; Yochum et al., 2012].
(iii) The critical shear stress can be seen as a key link in the feedback system between bed gradient, roughness (morphology), flow conditions and bedload transport. Empirical critical shear stress approaches, combined with a flow resistance equation, resulted in estimates of $D_{84}$ that follow the observed trend of increasing $D_{84}$ with bed slope (Figure 9). These field observations demonstrate that these critical shear stress approaches are valid for very steep channel slopes.

ACKNOWLEDGEMENTS

This study was supported by the CCES project APUNCH of the ETH domain. We thank K. Steiner, N. Zogg, M. Schneider and A. Beer for field assistance. We thank R. Hodge for providing TLS data analysis support, and K. Bunte for providing photos and pebble count grain size distributions.

APPENDIX A

Below, we present the derivation of the critical grain size as a function of bed slope, $D_{84}=f(S)$. It is based on combining the approaches of Camenen [2012] (Eq. 7) and Lamb [2008] (Eq. 8), which both define critical shear stress as a function of bed slope $S$, $\tau^*_c=f(S)$, together with the flow resistance equation of Ferguson (Eq. A1),

$$\frac{v}{v^*} = \frac{a_1 a_2 d}{D_{84}} \sqrt{a_1^2 + a_2^2 \frac{d}{D_{84}^{5/3}}} \quad (A1)$$

where $d$ is flow depth, $a_1=9.5$ and $a_2=3.9$ (these coefficients best fit the flow measurements at the Riedbach, see Figure 7a), $v$ is mean flow velocity, and $v^*$ is shear velocity. Replacing $v^*=(gdS)^{0.5}$ and $v=q/d$ based on the continuity equation, Eq. A1 can be converted to:

$$q = \frac{dg^{0.5} S^{0.5} a_1 a_2 d^{1.5}}{\sqrt{a_1 + a_2 \left( \frac{d}{D_{84}} \right)^{5/3}}} \quad (A2)$$

Squaring and simplifying Eq. A2 results in:

$$q^2 = \frac{g S a_1^2 a_2^2 d^2}{\left( a_1 + a_2 \frac{d}{D_{84}} \right)^{5/3}} \quad (A3)$$

The critical dimensionless shear stress is considered as a function of bed slope $f(S)$ as given in Eq. A4. Solving for flow depth $d$ and considering $w=D_{84}/D_{50}$ results in Eq. A5.

$$\tau^*_c = \frac{dS}{(s-1)D_{50}} = f(S) \quad (A4)$$
Chapter II

\[ d = D_{s0} \frac{(s-1)f(S)}{S} = \frac{D_{sA}}{w} \frac{(s-1)f(S)}{S} \quad (A5) \]

Now Eq. A5 can be inserted into Eq. A3 resulting in:

\[ \frac{q^2}{a_1 + a_2 \left( \frac{s-1}{w^{5/3}} \right) \frac{[f(s)]^{5/3}}{S^{5/3}}} = \frac{D_{sA}}{w} \frac{(s-1)f(S)}{S^{1/3}} \quad (A6) \]

Finally, Eq. A6 can be solved for \( D_{sA} \):

\[ D_{sA} = w^{5/3} q^{2/3} \left[ \frac{a_1 + a_2}{w^{5/3}} \left( s-1 \right) \frac{[f(S)]^{5/3}}{S^{5/3}} \right]^{1/3} \quad (A7) \]

REFERENCES


Fehr, R. (1987), A method for sampling very coarse sediments in order to reduce scale effects in movable bed models, paper presented at Proceedings of IAHR Symposium on Scale Effects in Moddelling Sediment Transport Phenomena, Toronto, Canada.


Chapter II


Leopold, L. B., R. A. Bagnold, M. G. Wolman, and L. Brush (1960), Flow resistance in sinuous or irregular channels, paper presented at The Physics of Sediment Transport by Wind and Water, ASCE.


Schneeg, P. A. (2003), A new field fluorometer for multi-tracer tests and turbidity measurement applied to hydrogeological problems, in *8th International Congress of the Brazilian Geophysical Society*, edited, Brazilian Geophysical Society, Rio de Janeiro, Brazil.


Weingartner, R., F. Hauser, A. Hermann, M. Probst, T. Reist, B. Schädler, and J. Schwanbeck (2014), Hydrologischer Atlas der Schweiz, edited, Geographisches Institut, Universität Bern; Bundesamt für Umwelt BAFU.


Chapter II


CHAPTER III

Applicability of bedload transport models for mixed size sediments in steep streams considering macro-roughness

Johannes M. Schneider, Dieter Rickenmann, Jens M. Turowski, Kristin Bunte, James W. Kirchner

Abstract - In steep mountain streams, macro-roughness elements typically increase flow-energy dissipation, and increase the threshold for initiation of motion compared to lower-gradient channels. Both effects reduce the flow energy available for bedload transport. Bedload transport models typically take account of these effects either by reducing the acting bed shear stress by the amount attributed to macro-roughness elements or by increasing the critical parameters for particle entrainment. Here we evaluate bedload transport models for mixed size sediments and models based on a median grain size using a large field dataset of fractional bedload transport rates, including 33 streams with channel bed gradients ranging from 0.05% to 11%. We derive the reference shear stress and bedload transport relations for using either the total boundary shear stress, or the reduced (termed “effective”) shear stress that accounts for flow resistance due to macro-roughness. If the reference shear stress is derived from the total boundary shear stress, it is closely related to channel slope, whereas if it is derived from the effective shear stress, it is almost constant over channel slope. The performance of bedload transport models is generally comparable either when using the total shear stress and a channel slope-related reference shear stress, or when using the effective shear stress and a constant reference shear stress. However, whereas the increase in the dimensionless bedload transport rate with hydraulic forcing is significantly steeper for the total stress approach, this increase is similar to the commonly used fractional Wilcock and Crowe transport model for the effective stress approach.

1. Introduction

Bedload transport is an important natural hazard in mountain regions, as well as a key process in landform evolution. Accurately predicting bedload transport, however, is difficult due to the complex interactions between the channel bed morphology and the hydraulic conditions.

The driving force for bedload transport is the hydraulic stress, which in bedload transport equations is most commonly expressed using discharge, stream power or shear stress. Shear stress is used throughout this study, but the presented concepts would be similar for discharge and stream power. In steady uniform flow, the dimensional shear stress \( \tau \) and the dimensionless shear stress \( \tau^* \) are related to the channel bed slope \( S \) [m/m] and flow depth or hydraulic radius \( h \) [m] according to

\[
\tau = \rho ghS, \quad \tau^* = \frac{\tau}{(\rho_s - \rho)gD} \tag{1a,b}
\]
Mixed-size sediments, macro-roughness and bedload transport

Where $g$ is gravitational acceleration [m/s$^2$], $\rho$ is water density [kg/m$^3$], $\rho_s$ is sediment density [kg/m$^3$] and $D$ is the diameter of a particle or grain size fraction [m]. In this study, we simply use the term ‘shear stress’ when referring to $\tau^*$ and the term ‘dimensional shear stress’ when we refer to $\tau$.

In mountain streams, bedload transport is not only influenced by parameters specified in Equation (1b), but also by the greater form roughness and structural bed stability compared to lower-gradient streams. Form roughness may be higher in steep streams due to (i) the channel form including pool-riffle to plane-bed, step-pool and cascade morphologies [Montgomery and Buffington, 1997]; (ii) an abundance of immobile boulders [e.g. Yager et al., 2007, Nitsche et al., 2011], (iii) a variable channel width and (iv) an increased number of bedrock constrictions and large woody debris jams [e.g. Wohl, 2000]. All these components contribute to what is commonly termed ‘form roughness’ or ‘macro-roughness’, which significantly increases total flow resistance and reduces the fraction of the flow energy available for sediment transport [e.g. Rickenmann and Recking, 2011; Nitsche et al., 2011]. If these effects are not taken into account, predicted bedload transport rates overestimate observed bedload transport by orders of magnitudes [e.g. Bathurst et al., 1987; Chiari and Rickenmann, 2011; Lenzi et al., 1999; Rickenmann, 2001; Rickenmann, 2012; Yager et al., 2007; Yager et al., 2012a].

Bedload transport in steep streams is also affected by their greater structural bed stability and therefore erosion resistance [Bunte et al., 2013]. Empirical evidence shows that the energy needed to initiate sediment motion and transport sediment is increased in steep streams. Possible explanations for the increased energy demand for sediment motion are given by Lamb et al. [2008] as ‘variable friction angles, grain emergence, flow aeration, changes to local flow velocity and turbulent fluctuations’, in addition to increased drag from macro-roughness. Two concepts exist for describing the critical conditions for initiation of particle motion in bedload transport calculations [see e.g. Bathurst, 2013; Wilcock, 1988]: the flow competence method, in which the critical condition is defined by the maximum particle size moved at a given flow with a related ‘critical’ shear stress [Shields, 1936], and the reference transport method, in which the critical condition is defined by a ‘reference’ shear stress which corresponds to an arbitrarily defined small bedload transport rate [Parker et al., 1982]. An increase in the critical shear stress with increasing channel slope has been shown for both the critical Shields stress [Bunte et al., 2013; Lamb et al., 2008; Prancevic et al., 2014; Recking, 2009; Shvidchenko et al., 2001] and the reference shear stress [Mueller et al., 2005]. These bed-slope-dependent critical shear stress values have all been derived using the total boundary shear stress, which does not account for energy dissipation due to macro-roughness elements. Other studies have identified relationships for the reference shear stress as a function of sand content [Wilcock and Crowe, 2003], the geometric standard deviation of the bed surface grain size distribution [Gaumann et al., 2009] or the degree of armouring [Efthymiou, 2012].

The critical conditions for initiation of motion of individual particle size classes are important for predicting fractional as well as total transport rates. Particle mobility is typically related to absolute particle size, and larger, heavier particles are assumed to be more difficult to mobilize than smaller, lighter particles [Buffington and Montgomery, 1997; Shields, 1936]. But in beds with mixed grain sizes, as are typical for steep mountain streams, particle mobility depends not only on the absolute grain size but also on the relative grain size.
Smaller particles are hidden behind larger pieces of gravel, and are therefore protected from the flow, while larger particles are more exposed to the flow [e.g., Egiazaroff, 1965; Einstein, 1950; Parker, 2008; Parker and Klingeman, 1982; Parker et al., 1982; Wiberg and Smith, 1987]. These effects are known as hiding and exposure, and they can modulate the relative mobility of different grain size fractions. If hiding and exposure effects exactly cancel out the effects of particle mass, the probability of entrainment for each grain is independent of grain size at a given hydraulic stress, resulting in a condition termed ‘equal mobility’. The critical conditions for initiation of motion of individual particle size classes are typically described in terms of the fractional reference shear stress $\tau_{ri}^*$ using a so-called hiding function. Within a hiding function, the ratio between $\tau_{ri}^*$ and $\tau_{rD50}^*$, the reference shear stress for the median grain size, is related to the ratio $D_i/D_{50}$ by the hiding exponent $b$ (Equation (2)).

$$\frac{\tau_{ri}^*}{\tau_{rD50}^*} = \left(\frac{D_i}{D_{50}}\right)^b$$

The exponent $b$ varies from 0 for full size selectivity in the absence of relative size effects to -1 for equal mobility [e.g., Ashida and Michiue, 1973; Andrews et al., 1987; Bathurst, 2013; Parker 1990; Parker and Klingeman, 1982; Parker et al., 1982; Powell et al., 2001; Recking, 2009; Wilcock and Crowe, 2003].

Although most proposed bedload transport functions are based on physical consideration of the transport process, they also depend on empirical fits to data from flume studies or field observations. Important elements of this fitting procedure are illustrated with the help of Figure (1), which represents synthetic bedload transport data in terms of the dimensionless transport rate $W^*$ and the bed shear stress ratio $\tau^*/\tau_r^*$ (Equation (3)).

$$W^* = \frac{Rgq_{bVol}}{u^*^3}$$

Where $R = \rho_s/\rho_s - 1$, $q_{bVol}$ is the volumetric bedload transport rate per unit width [m$^3$/sm], and $u^*$ is the shear velocity ($u^* = (\tau/\rho)^{0.5}$).

Three main elements can be important in the fitting procedure: (i) calculation of the reduced or ‘effective’ shear stress $\tau^*$, (ii) determination of reference shear stress $\tau_r^*$, and (iii) choice of the transport function including its pre-factor. One main challenge in the application of a bedload transport formula is calculating the flow energy available for bedload transport, i.e. an effective shear stress $\tau^*$ (which in Figure (1) will move the gray experimental data horizontally to the left). One of the earliest approaches to account for flow resistance due to macro-roughness is to partition total flow resistance into grain and from resistance, as originally proposed by Meyer-Peter and Mueller [1948]. Similar approaches to partitioning between grain and form resistance, or between base-level and macro-roughness, were later suggested by others [e.g., Carson, 1987; Carson and Griffiths, 1987; Gomez and Church, 1989; Millar and Quick, 1994; Millar, 1999; Palt, 2001; Parker and Peterson, 1980; Rickenmann and Recking, 2011; Rickenmann et al., 2006].
Mixed-size sediments, macro-roughness and bedload transport

Figure 1: Illustration of a similarity collapse for bedload transport model development. Varying the effective shear stress $\tau^*$ and/or the reference shear stress $\tau^*_r$ allows the scattered data points (gray dots) to be moved left or right. Ideally it is possible to collapse the adjusted datapoints to represent a well-defined relation (blue dots). The mathematical bedload transport equation (black line) determines the rate of change of transport intensity with increasing shear stress, and the pre-factor determines the vertical position in the diagram. Note: data points in this illustration are randomly generated. The shape of the example bedload relation corresponds to the Wilcock and Crowe [2003] equation.

However, applications of a reduced effective shear stress $\tau^*$ to bedload transport calculations, and comparisons with field data, remain limited [Chiari and Rickenmann, 2011; Gaueuman et al., 2009; Nitsche et al., 2011; 2012; Wilcock, 2001; Wilcock et al., 2009]. Another approach to collapsing the experimental data onto a single function is based on the reference shear stress approach of Parker et al. [1982]. In case of fractional transport models, which take relative grain size effects such as hiding and exposure into account, the collapse of fractional transport rates is based on the fractional reference shear stress of each grain size fraction ($\tau^*_r$); this approach is described in more detail below in section 2.3.2.

The reference shear stress assumes values comparable to the critical shear stress $\tau^*_c$ defined for threshold-based bedload transport equations. For example, Wilcock et al. [2009] demonstrated that for the Meyer-Peter and Mueller [1948] equation $\tau^*_c = 0.996 \tau^*_r$. The choice of an appropriate values for $\tau^*_r$ (or $\tau^*_c$) affects the horizontal position of the experimental data in Figure (1).

The third important element refers to the choice of the transport function and its pre-factor. Concerning the Meyer-Peter and Mueller [1948] transport equation, both Hunziker and Jaeggi [2002] and Wong and Parker [2006] suggested to reduce the pre-factor from a value of 8 in the original equation to a value of about 5, based on a reanalysis of the experimental data. For a reference-shear-stress-based transport equation, such as that of Wilcock and Crowe [2003], the dimensionless transport rate $W^*$ increases as a power-law function of the bed shear stress ratio $\tau^*/\tau^*_r$, with an exponent $m = 7.5$ for $\tau^*/\tau^*_r$ values smaller than about 1.3. Recking [2010] developed a reference-shear-stress-based transport equation, and for $\tau^*/\tau^*_r$ values smaller than about 1, he proposed a power law exponent $m = 12.9$ based on flume data.
with uniform bed material, whereas he modified the exponent to $m = 6.5$ with the inclusion of field data from gravel-bed streams, which obviously represent a non-uniform bed material size distribution. For this latter transport equation, Recking [2010] furthermore introduced a correction function for the pre-factor, which depends on channel slope and on the ratio $D_{84}/D_{50}$; however, he did not explicitly account for a reduced effective shear stress $\tau^*$. In terms of Figure (1), the choice of transport function will affect mainly the steepness of the fitted transport relation while its pre-factor will affect the vertical position relative to the experimental data. In summary, any empirical fitting of a bedload transport function implies quantification of the three elements (i, ii, and iii) discussed above.

Although it is obvious that the flow energy available for sediment transport at steep slopes is limited due to macro-roughness, and although it could be shown in flume experiments that the energy needed for particle entrainment is increased on steep beds [Lamb et al., 2008; Mueller et al., 2005; Prancevic et al., 2014; Recking, 2009], it remains almost impossible to distinguish between these two effects, macro-roughness and slope, in field data. For bedload transport prediction at steep slopes, one therefore commonly chooses to reduce the total boundary shear stress [e.g., Egashira and Ashida, 1991; Nitsche et al., 2011; Pagliara and Chiavaccini, 2006; Rickemann 2012; Whittaker et al., 1988; Yager et al., 2007; Yager et al., 2012b] or increase the critical threshold [e.g., Bunte et al., 2013; Lamb et al., 2008; Mueller et al., 2005; Prancevic et al., 2014] to account for the reduced available flow energy. However, it has not been shown how the choice of using one or the other approach affects bedload transport prediction.

In this study we present a high-quality field dataset of fractional bedload transport data for grain sizes larger than 2-10 mm (depending on field site) covering a wide range of channel slopes and stream characteristics. Based on the field data we analyze how using either total or effective shear stress in the adjusted reference shear stress affects the similarity collapse between the dimensionless transport rate $W^*$ and the bed shear stress ratio $\tau^*/\tau^*_r$. We further examine how different adjustments of the bedload transport function, which are based either on a fractional- or a total bedload transport data, affect bedload transport predictions. We demonstrate that by using a reduced-effective-shear-stress approach, the adjusted reference shear stress values fall in a similar range as for flatter channel gradients, and the shape of the transport function for small to medium transport rates is similar to the one developed by Wilcock and Crowe [2003] based on flume experiments. We discuss our results with the findings of other studies and attempt to assess the relative importance of the different adjustment possibilities.

2. Field data and analysis

The field sites used to investigate bedload transport cover a wide range of channel slopes, flow discharge rates and grain-sizes (Table (1)). Channel slopes range from about 0.05% to 11%, bankfull flows range from 0.3 to 114 m$^3$/s, and the median grain size of the surface grain size distribution $D_{50surf}$ ranges from 0.004 to 0.2 m. The selected streams are compiled from the bedload trap data provided by Bunte et al. [2004, 2008, 2010, 2014] (denoted Bunte in this study) including 10 US streams in Colorado, Wyoming and Oregon, as well as data from Oak Creek, Oregon, US [Milhous, 1973], and data from the Swiss streams Erlenbach.
Mixed-size sediments, macro-roughness and bedload transport

(denoted EB) and Riedbach (denoted RB). In addition, the dataset was complemented by literature data found in King et al. [2004] and Williams and Rosgen [1989], denoted as King and WR, respectively. The King dataset has been analyzed in detail in many previous publications [e.g., Barry, 2004; Barry et al., 2008; Mueller et al., 2005; Muskatirovic, 2008; Recking, 2010, Whiting et al., 1999].

2.1. Stream bed grain-size distributions

For the Bunte streams as well as for the two Swiss streams Erlenbach and Riedbach, bed material surface grain-size distributions (GSDs) have been determined by grid-by-number pebble counts using a variant of the Wolman [1954] procedure, in which a sampling frame indicates particles to be included in the sample, and particle sizes are measured with a template [Bunte and Abt, 2001a; b; Bunte et al., 2009]. Sand-size particles are lumped into the <2 mm category. Bed material subsurface GSDs were determined by sieve analysis of large volumetric samples taken using a three-sided plywood shield [Bunte and Abt, 2001b].

For the King dataset and the Colorado streams of the WR dataset, bed surface diameters were obtained by the Wolman [1954] pebble count method, where sand particles mostly are also lumped into a category of <2 mm, and subsurface samples were collected in open-ended, 0.5 m diameter barrels. Measurement techniques for the Alaska streams of the WR dataset are not fully reported [see also Knott and Lipscomb, 1984].

The grain-size distributions of all streams, which were originally reported in various scales, were converted to the full phi scale > 2 mm (2, 4, …1064 mm) for simplified computing and analysis. The minimum grain size of 2 mm was chosen because it refers to the typical minimum grain size of the grain-size distributions derived from pebble counts as applied for most of the streams during this study. For the King and WR dataset, the $D_{50}$Surf and $D_{84}$Surf grain sizes were derived from the converted grain-size distributions. This explains why the characteristic grain sizes used in this study occasionally deviate from the values reported in other studies.

2.2. Bedload transport data and measurement techniques

Fractional bedload transport rates were determined for the Bunte dataset and the Riedbach data [Schmid, 2011] using portable bedload traps. Bedload traps consist of an aluminum frame (0.3 x 0.2 m opening) with a 1-1.6 m long nylon net attached at its downstream end to capture bedload material [Bunte et al., 2004; Bunte et al., 2007]. The nylon net had a mesh width of 4 mm (in the Riedbach the mesh width was 6 mm; the issue of how we treat different minimum measured grain sizes will be discussed at the end of this section). The traps are placed onto ground plates that are anchored in the streambed by metal stakes and then strapped to the stakes. The fixed anchors and large nets facilitate long sampling times (typically an hour) and large sample volumes (up to 20 liters), thus improving the chances of proportionally sampling the coarse bedload fractions that move less frequently.

The Oak Creek (USA) bedload transport measurements were taken using a vortex sediment sampler installed over the full width of the channel [Milhous, 1973]. The Oak Creek sediment transport data have been used in several publications for deriving bedload transport equations [e.g., Milhous, 1973; Parker, 1990; Parker et al., 1982].
### Table 1: Stream characteristics a)

<table>
<thead>
<tr>
<th>Source</th>
<th>Stream b)</th>
<th>State, Country</th>
<th>( S ) [m/m]</th>
<th>( A ) [m²]</th>
<th>( Q ) [cfs]</th>
<th>( D ) [m]</th>
<th>( D ) [m]</th>
<th>Number of values</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bunte</td>
<td>Helley and Smith</td>
<td>Alaska, USA</td>
<td>0.013</td>
<td>16369</td>
<td>0.02</td>
<td>0.05</td>
<td>0.10</td>
<td>39</td>
</tr>
<tr>
<td>Bunte</td>
<td>Talkeetna R.</td>
<td>Alaska, USA</td>
<td>0.013</td>
<td>16369</td>
<td>0.02</td>
<td>0.05</td>
<td>0.10</td>
<td>39</td>
</tr>
<tr>
<td>Bunte</td>
<td>Lo. NF Cabin Cr.</td>
<td>Colorado, USA</td>
<td>0.0007</td>
<td>16369</td>
<td>0.02</td>
<td>0.05</td>
<td>0.10</td>
<td>39</td>
</tr>
<tr>
<td>Bunte</td>
<td>Buffalo Cr.</td>
<td>Colorado, USA</td>
<td>0.0007</td>
<td>16369</td>
<td>0.02</td>
<td>0.05</td>
<td>0.10</td>
<td>39</td>
</tr>
<tr>
<td>Bunte</td>
<td>Horse Cr.</td>
<td>Colorado, USA</td>
<td>0.0007</td>
<td>16369</td>
<td>0.02</td>
<td>0.05</td>
<td>0.10</td>
<td>39</td>
</tr>
</tbody>
</table>

a) \( S \): Channel slope; \( A \): Basin area; \( Q \): bank full flow or return period of 1.5 years respectively (\(^\dagger\)); b) Original stream descriptions occasionally abbreviated (Cr.=Creek; E=East; N=North; F=Fork; W=Wood; M=Main/Middle; L=Left; Lo=lower; bl.=below; nr.=Near; R=River; S=South). c) MBB: Moving bedload basket system [Rickenmann et al., 2012]; BLT: Bedload trap [Bunte et al., 2004]; HS: Helley and Smith [1971] sampler with a 76 mm opening; Vort.: Vortex tube [Klingeman, 1979; Millhous, 1973 (MH)].
Mixed-size sediments, macro-roughness and bedload transport

At the Erlenbach, an automatic moving bedload basket (MBB) system was installed in 2009, which provides fractional bedload transport rates together with discharge measurements [Rickenmann et al., 2012]. Movable, 1 m³ metal baskets with 10 mm mesh size are mounted on a rail at the downstream side of a large check dam above a retention basin. The baskets can be moved automatically into the flow and the samples are subsequently sieved to determine grain size fractions. Sampling times varied according to transport rates and could be as short as a few seconds at high discharges in order to not exceed a sample mass of around 200 kg.

Our database is complemented by two published data sets compiled by Williams and Rosgen [1989] and King et al. [2004] to extend the range of bed gradients and grain sizes. These data were collected with Helley and Smith [1971] (HS) pressure difference samplers. The HS measurement technique was validated for sand and fine gravel by Emmett [1980], who found a similarity between sand and gravel bedload transport rates collected with a HS sampler on a concrete sill and those from a conveyor-belt sampler. The HS samplers used for the streams have a 76.2 by 76.2 mm square orifice and a 0.0625-0.25 mm mesh size on the collecting bag. Typically, the handheld sampler is set directly onto the streambed while the operator is either wading the stream or deploying the HS from a bridge.

Because our study is based on rather well-defined field-measured transport relations, we used only those WR and King data set streams that yielded a coefficient of determination $r^2 > 0.5$ for fitted bedload transport relations in the form of

$$q_b = \alpha Q^\beta$$

Here $q_b$ is the total bedload transport rate per unit channel width [kg/sm], $Q$ is the flow discharge [m³/s] and $\alpha$ and $\beta$ are empirical constants. We fitted Equation (4) to all available data using linear least-squares regression on log-transformed values; $r^2$ was also determined on the log-transformed values.

In general, the HS samples have larger sampling uncertainties compared to other measurement techniques described in this study. One reason is the small 76.2x76.2 mm² opening compared to the 300x200 mm² opening of the bedload traps; another reason is the shorter sampling time (typically around 20-180 sec) [for more details see Bunte et al., 2008]. Therefore, we decided to analyze the HS samples separately from the data derived from the other measurement techniques. In the following, the data by Bunte, the Riedbach, Erlenbach and Oak Creek data are denoted as the Main Dataset, while the dataset including the HS samples including the WR and King data is denoted as the HS Dataset.

2.2.1 Bedload grain-size distributions

The bedload grain-size distributions (GSD) were converted to the same full phi scale as the bed surface and subsurface GSDs. The smallest sampled bedload grain sizes vary among datasets depending on the measurement technique and range from 0.0625 mm for the WR dataset to 10 mm for the Erlenbach. We truncated the King and WR bedload grain-size distributions at 2 mm to exclude near-bed suspended fines. Also for the following dimensionless analysis of the fractional transport rates we relate fractional transport rates to surface grain size fractions, which are only available/valid for size fractions larger than 2 mm (see Section (2.1.)). The following analysis relating fractional transport rates to the reference
shear stress and acting bed shear stress is not strongly affected by the variation in the minimum measured grain size (2 mm to 10 mm). For the analysis of total transport rates we define total bedload transport as larger than 4 mm. This corresponds to the minimum grain size measured by the bedload traps. The resulting uncertainties/biases in total transport rates for the Riedbach ($D > 6 \text{ mm}$) and the Erlenbach ($D > 10 \text{ mm}$) are neglected.

### 2.3. Flow hydraulics

Bedload transport is described throughout this study in relation to dimensionless shear stress $\tau^*$, which is proportional to flow depth (or hydraulic radius) and channel slope (Equation (1)). Moreover, flow depth is often unmeasured in stream studies or may have large sampling uncertainties. Recking [2010] argued that the predictive quality of bedload equations is improved when discharge is used instead of flow depth. Therefore we back-computed the hydraulic radius for the study streams from the measured flow discharge and average cross-sectional profile. We iteratively computed flow depth and width, hydraulic radius and cross-sectional area for each measured discharge and reach-averaged velocity, using the variable power equation (VPE) of Ferguson [2007]:

$$\frac{U}{v^*} = \frac{a_1 a_2 \left( \frac{rh}{D_{84Surf}} \right)}{\sqrt{a_1^2 + a_2^2 \left( \frac{rh}{D_{84Surf}} \right)^{5/3}}}$$  \hspace{1cm} (5)

where $U$ is flow velocity [m/s], $v^*$ is dimensionless shear velocity ($v^* = (grhS)^{0.5}$), $rh/D_{84}$ is the relative flow depth, $a_1 = 6.5$ and $a_2 = 2.5$ (see Rickenman and Recking, [2011]). Rickenman and Recking [2011] evaluated several flow resistance equations based on a large field dataset of flow velocity measurements in gravel-bed rivers and found that the VPE of Ferguson [2007] generally provided the best performance (see also Nitsche et al., [2011]). Equation (5) implicitly accounts for large roughness elements and was found to also give good predictions of total flow resistance for flow conditions with small relative flow depths ($rh/D_{84}$) in steep streams. The iterative calculation of the hydraulic parameters using Equation (5) was performed on average cross-sections derived from surveyed cross-sections upstream of the sampling location for the Riedbach, the Erlenbach and the King et al. [2004] datasets. For the US bedload trap study, the cross-sections surveyed at or near the bedload trap location were used for computing the hydraulic parameters. For the Williams and Rosgen [1989] dataset, a rectangular cross-section was estimated from the given stream width.

#### 2.3.1 Determination of effective shear stress $\tau'$

Two approaches are used in this study to estimate the increased flow resistance in steep streams due to macro-roughness elements (boulders, bedrock, and woody debris). The first approach (Rickenmann and Recking [2011], denoted here RR2011) is based on computing dimensional shear stress using a reduced energy slope $S_{red}$ (Equation (6)) instead of the actual channel slope [Chiari and Rickenmann, 2011; Chiari et al., 2010; Nitsche et al., 2011; Rickenmann and Recking, 2011; Rickenmann, 2012]. The concept of a reduced energy slope is based on flow-resistance partitioning between a base level flow resistance ($f_0$) and the total resistance ($f_{tot}$), which includes additional resistance due to large roughness elements at
Mixed-size sediments, macro-roughness and bedload transport

relatively small flows [for more details see Rickenmann and Recking, 2011; Rickenmann, 2012]:

\[ S_{red} = S \left( \frac{f_0}{f_{tot}} \right)^e \]  \hspace{1cm} (6a)

\[ \tau' = \rho gh S_{red} \]  \hspace{1cm} (6b)

In related studies an exponent of \( e = 1.5 \) often resulted in a considerable improvement in predicted bedload transport, and the same value of \( e = 1.5 \) was used in this study. This approach is similar to flow resistance partitioning between grain and form resistance as proposed in earlier studies [Millar and Quick, 1994; Millar, 1999].

We also used the approach of Wilcock et al. [2009] (denoted here WC2009) to account for form resistance or macro-roughness, which generally corresponds to the approach earlier suggested in Wilcock [2001]. The Wilcock et al. [2009] equation determines grain resistance using a Manning-Strickler relationship and defines the reduced (effective) dimensional shear stress \( \tau' \) via the \( D_{65} \) of the streambed surface (in mm):

\[ \tau' = 17 (SD_{65})^{0.25} U^{1.5} \]  \hspace{1cm} (7)

where \( U \) is the flow velocity in m/s. Dimensional shear stress values \( \tau' \) from Equation (7) are very similar to those obtained using Equation (6). For the sake of a simplified presentation, the results based on Equation (7) are presented only in the supporting information. Both Equations (6) and (7), are referred to dimensional forms of the shear stress and can be non-dimensionalized using Equation (1b).

2.3.2 Derivation of the reference shear stress

The adjustment of the fractional reference shear stress, used to collapse the fractional dimensionless transport rates (both within individual streams and between different streams), was determined following the reference approach of Parker et al. [1982] and Wilcock and Crowe [2003]. The dimensionless reference transport rate used in this study \( W^* \) for the \( i \)th size fraction was set to 0.002. The dimensionless transport rate \( W^*_i \) is defined by Equation (8) as

\[ W^*_i = \frac{Rgq_{biVol}}{F_i u^*^3} \]  \hspace{1cm} (8)

where \( F_i \) is the proportion of grain size \( i \) on the bed surface, \( q_{biVol} \) the volumetric fractional bedload transport rate per unit width [m\(^3\)/sm] and \( u^* \) is the shear velocity \( (u^* = (\tau/\rho)^{0.5}) \). The reference approach of Parker et al. [1982] includes the following four steps:

- Fitting power functions \( (W^*_i = a_i \tau_i^m_i) \) to \( W^*_i \) and \( \tau_i^* \) for each individual grain size fraction \( i \) (shown for the East Dallas Creek as an example in Figure (2a)).
- Fitting functions \( (W^*_i = a_i \tau_i^{m_{i50}}) \) for each grain size fraction with \( m_{i50} \) as the median of the exponents \( m_i \) derived in the previous step (Figure (2a)).
Chapter III

- Estimating $\tau_{*ri}^{*}$ by selecting the value of $\tau_{*i}$ corresponding to a reference transport rate of $W_{ri}^{*} = 0.002$.

- Fitting a relation of $\tau_{*ri}^{*}$ against $D_{iSurf}/D_{50Surf}$, thus determining the reference shear stress $\tau_{riD50}^{*}$ and the hiding exponent $b$ (Figure 2b)).

Based on the hiding function approach it is possible to collapse fractional transport rates for a given channel reach onto a more-or-less single curve (Figure (2c)). The resulting fractional transport power-law relationship is defined in this study by the pre-factor 0.002 and the median exponent $m_{50}$ (Figure (2c)).

In addition to the fractional reference shear stress, the reference shear stress $\tau_{*riD50Surf}^{*}$ was also determined from total transport rates for $D>4$ mm by fitting power functions to $W_{tot}^{*}$ vs. $\tau_{*D50Surf}^{*}$ and identifying the intercept of the power law with $W_{tot}^{*} = 0.002$ (see also Mueller et al. [2005]). The 4 mm criterion for $W_{tot}^{*}$ was chosen because it is the minimum grain size sampled with bedload traps, except at the Riedbach (6 mm) and the Erlenbach (10 mm). For those total transport computations, all HS Samples were truncated at 4 mm as well.

Figure 2: Reference approach [Parker et al., 1982] applied to East Dallas Creek (Bunte Dataset). (a) Power functions fitted to plots of fractional dimensionless bedload transport rates vs. dimensional shear stress for 1-phi size classes from 4 mm (o) to 32 mm (*), with free exponents $m$ (grey lines) and a fixed averaged exponent $m_{50}=13.7$ (black lines). (b) Hiding function in the form of Equation (2) with a constant exponent $b$ (black line) and a variable exponent according to Wilcock and Crowe [2003] (grey dotted line, Equation (A4)). (c) Collapse of fractional transport rates. Horizontal lines in a) and c) represent the dimensionless transport rate $W_{*}=0.002$. 

53
3. Results

3.1. Reference shear stress ($\tau_{rD_{50}}$)

3.1.1 $\tau_{rD_{50}}$ - Derived from fractional bedload transport rates

The reference shear stress was derived from fractional transport rates for each stream using the total boundary shear stress ($\tau_{rD_{50}}$), as well as the reduced effective shear stress ($\tau_{*rD_{50}}$) according to Equation (6) [Rickenmann and Recking, 2011] and Equation (7) [Wilcock et al., 2009] (Table (2), Table S1). $\tau_{rD_{50}}$ derived from the total boundary shear stress was related to various stream characteristics and compared with results from other studies (Figure (3)): channel slope [Bunte et al., 2013; Lamb et al., 2008; Mueller et al., 2005], sand content [Wilcock and Crowe, 2003], degree of armouring shown by the $D_{50Surf}/D_{50Sub}$ ratio [Efthymiou, 2012], and two measures of the width of the grain size distribution: its geometric standard deviation $\sigma_{sg}$ (Equation (9)) [Gaeuman et al., 2009], and the $D_{84Surf}/D_{30Surf}$ ratio.

All power laws were only fitted to the Main Dataset. The reference shear stress $\tau_{rD_{50}}$ is related to channel slope (Figure (3a)) according to

$$\tau_{rD_{50}} = 0.44S^{0.44} \text{ for } S > 0.01; \ r^2 = 0.63$$  \hspace{1cm} (10)

which follows a similar pattern to that reported by Mueller et al. [2005] for the King data. The relationship between $\tau_{rD_{50}}$ and channel slope also resembles the critical shear stress relations of Bunte et al. [2013] and Lamb et al. [2008]. Note that the Bunte et al. [2013] relation for the critical shear stress was derived from the same bedload transport data as used here, including the Oak Creek data; however, the calculation procedure corresponds to the flow competence approach (See Section (1)). The values of the HS Dataset generally produce $\tau_{rD_{50}}$ values smaller by up to one order of magnitude than those of the Main Dataset, but $\tau_{rD_{50}}$ values in both datasets exhibit similar trends with channel slope (Figure (3a)). Furthermore, $\tau_{rD_{50}}$ is related to the $D_{84Surf}/D_{30Surf}$ ratio and $\sigma_{sg}$ (Figure (3e, 3f)). However, $\tau_{rD_{50}}$ is positively correlated to $\sigma_{sg}$, whereas Gaeuman et al. [2009] described a negative trend. Note that the Gaeuman et al. [2009] relation was based on a reduced effective shear stress according to Wilcock [2001], similar to the approach for shear stress reduction used in Equation (7) in this study.

A correlation of $\tau_{rD_{50}}$ with sand content (the $D<2$ mm fraction of the original subsurface grain size distributions, before truncation) as derived by Wilcock and Crowe [2003] in flume experiments could not be observed for the field data. Neither could we identify a correlation of $\tau_{rD_{50}}$ with the $D_{50Surf}/D_{50Sub}$ ratio as proposed from laboratory data by Efthymiou [2012].

For the reduced effective shear stress (Figure (4); see also Figure (S1)), the strong positive correlations of $\tau_{rD_{50}}$ with channel slope $S$, the $D_{84Surf}/D_{30Surf}$ ratio and $\sigma_{sg}$ (Figure (3a, 3e, 3f)) almost vanish. The $\tau_{rD_{50}}$ relation with channel slope in the Main Dataset (Figure (4a)) is highly scattered but tends to be negative; however, this negative trend is not supported by the independent HS data (see also Figure (5b)). There is no relationship between $\tau_{*rD_{50}}$ and $\sigma_{sg}$, and the values of $\tau_{*rD_{50}}$ are in a similar range as values reported by Gaeuman et al. [2009].
### Table 2: Hiding function ($\tau^*_j$) and dimensionless fractional bedload rating curves ($W^*_j$) derived from total boundary shear stress $\tau^*$ and the reduced effective shear stress $\tau^*_i$ according to Rickenmann and Recking [2011] (RR2011).

<table>
<thead>
<tr>
<th>Stream</th>
<th>$\tau^*_{iD50}$</th>
<th>$b$</th>
<th>$r^j$</th>
<th>$m_{iD50}$</th>
<th>$\tau^*_{iD50}$</th>
<th>$b$</th>
<th>$r^j$</th>
<th>$m_{iD50}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Erlenbach</td>
<td>0.19</td>
<td>-0.95</td>
<td>0.993</td>
<td>7.29</td>
<td>0.01</td>
<td>-0.86</td>
<td>0.941</td>
<td>2.79</td>
</tr>
<tr>
<td>Riedbach</td>
<td>0.11</td>
<td>-0.94</td>
<td>0.997</td>
<td>17.22</td>
<td>0.05</td>
<td>-0.88</td>
<td>0.982</td>
<td>8.74</td>
</tr>
<tr>
<td>St. Louis Cr.</td>
<td>0.09</td>
<td>-0.98</td>
<td>0.998</td>
<td>7.78</td>
<td>0.07</td>
<td>-0.95</td>
<td>0.993</td>
<td>3.84</td>
</tr>
<tr>
<td>Cherry Cr.</td>
<td>0.12</td>
<td>-0.96</td>
<td>1.000</td>
<td>21.45</td>
<td>0.07</td>
<td>-0.92</td>
<td>0.999</td>
<td>11.79</td>
</tr>
<tr>
<td>Litl.Granite Cr.A</td>
<td>0.07</td>
<td>-0.97</td>
<td>1.000</td>
<td>25.32</td>
<td>0.04</td>
<td>-0.95</td>
<td>1.000</td>
<td>14.89</td>
</tr>
<tr>
<td>E.St.Louis Cr.A$^{a)}$</td>
<td>0.14</td>
<td>-0.99</td>
<td>1.000</td>
<td>20.52</td>
<td>0.03</td>
<td>-0.97</td>
<td>0.999</td>
<td>9.13</td>
</tr>
<tr>
<td>E.St.Louis Cr.B</td>
<td>0.09</td>
<td>-0.97</td>
<td>0.999</td>
<td>10.65</td>
<td>0.05</td>
<td>-0.92</td>
<td>0.997</td>
<td>5.10</td>
</tr>
<tr>
<td>Halfmoon Cr.</td>
<td>0.11</td>
<td>-0.95</td>
<td>1.000</td>
<td>16.68</td>
<td>0.04</td>
<td>-0.91</td>
<td>0.999</td>
<td>7.94</td>
</tr>
<tr>
<td>Hayden Cr.</td>
<td>0.06</td>
<td>-0.85</td>
<td>0.999</td>
<td>13.67</td>
<td>0.04</td>
<td>-0.72</td>
<td>0.998</td>
<td>7.35</td>
</tr>
<tr>
<td>E.Dallas Cr.</td>
<td>0.07</td>
<td>-0.98</td>
<td>1.000</td>
<td>30.13</td>
<td>0.02</td>
<td>-0.97</td>
<td>0.999</td>
<td>13.19</td>
</tr>
<tr>
<td>Fool Cr.</td>
<td>0.15</td>
<td>-1.08</td>
<td>0.996</td>
<td>6.50</td>
<td>0.06</td>
<td>-1.15</td>
<td>0.986</td>
<td>2.69</td>
</tr>
<tr>
<td>NF Swan Cr.</td>
<td>0.05</td>
<td>-0.91</td>
<td>0.999</td>
<td>12.11</td>
<td>0.03</td>
<td>-0.83</td>
<td>0.995</td>
<td>6.57</td>
</tr>
<tr>
<td>Oak Cr.</td>
<td>0.04</td>
<td>-0.86</td>
<td>1.000</td>
<td>10.62</td>
<td>0.03</td>
<td>-0.75</td>
<td>0.999</td>
<td>5.65</td>
</tr>
<tr>
<td>Big W.R.nr.Ket.</td>
<td>0.04</td>
<td>-0.92</td>
<td>0.996</td>
<td>6.17</td>
<td>0.02</td>
<td>-0.86</td>
<td>0.988</td>
<td>3.45</td>
</tr>
<tr>
<td>Boise R.</td>
<td>0.02</td>
<td>-0.90</td>
<td>0.994</td>
<td>3.71</td>
<td>0.01</td>
<td>-0.87</td>
<td>0.987</td>
<td>2.93</td>
</tr>
<tr>
<td>Dollar Cr.</td>
<td>0.05</td>
<td>-0.92</td>
<td>0.996</td>
<td>4.65</td>
<td>0.02</td>
<td>-0.82</td>
<td>0.981</td>
<td>1.93</td>
</tr>
<tr>
<td>Herd Cr.</td>
<td>0.04</td>
<td>-0.12</td>
<td>0.955</td>
<td>3.40</td>
<td>0.03</td>
<td>-0.78</td>
<td>0.995</td>
<td>1.68</td>
</tr>
<tr>
<td>Lochsa R.</td>
<td>0.03</td>
<td>-0.84</td>
<td>0.996</td>
<td>6.42</td>
<td>0.02</td>
<td>-0.79</td>
<td>0.992</td>
<td>5.01</td>
</tr>
<tr>
<td>MF Red R.</td>
<td>0.04</td>
<td>-0.45</td>
<td>0.811</td>
<td>2.86</td>
<td>0.04</td>
<td>-0.11</td>
<td>0.082</td>
<td>1.76</td>
</tr>
<tr>
<td>MF Salmon R.</td>
<td>0.03</td>
<td>-0.93</td>
<td>0.997</td>
<td>11.84</td>
<td>0.02</td>
<td>-0.91</td>
<td>0.994</td>
<td>9.04</td>
</tr>
<tr>
<td>NFCl.Water R.</td>
<td>0.01</td>
<td>-1.06</td>
<td>0.994</td>
<td>5.79</td>
<td>0.01</td>
<td>-1.08</td>
<td>0.993</td>
<td>4.93</td>
</tr>
<tr>
<td>Rapid R.</td>
<td>0.06</td>
<td>-0.76</td>
<td>0.989</td>
<td>4.36</td>
<td>0.04</td>
<td>-0.48</td>
<td>0.881</td>
<td>2.01</td>
</tr>
<tr>
<td>Salmon Rbl.Ynk.F</td>
<td>0.03</td>
<td>-0.97</td>
<td>0.993</td>
<td>8.53</td>
<td>0.02</td>
<td>-0.96</td>
<td>0.986</td>
<td>6.10</td>
</tr>
<tr>
<td>Salmon R. nr. Obs.</td>
<td>0.04</td>
<td>-0.88</td>
<td>0.997</td>
<td>5.99</td>
<td>0.03</td>
<td>-0.83</td>
<td>0.993</td>
<td>5.13</td>
</tr>
<tr>
<td>Salmon R. nr. Shp.</td>
<td>0.02</td>
<td>-0.94</td>
<td>0.998</td>
<td>7.81</td>
<td>0.02</td>
<td>-0.93</td>
<td>0.997</td>
<td>6.83</td>
</tr>
<tr>
<td>Selway R.</td>
<td>0.02</td>
<td>-0.93</td>
<td>0.996</td>
<td>6.24</td>
<td>0.02</td>
<td>-0.92</td>
<td>0.994</td>
<td>5.10</td>
</tr>
<tr>
<td>SF Payette R.</td>
<td>0.02</td>
<td>-0.89</td>
<td>0.897</td>
<td>1.97</td>
<td>0.01</td>
<td>-0.82</td>
<td>0.731</td>
<td>1.18</td>
</tr>
<tr>
<td>Squaw Cr.USGS</td>
<td>0.04</td>
<td>-0.86</td>
<td>0.983</td>
<td>3.73</td>
<td>0.02</td>
<td>-0.76</td>
<td>0.937</td>
<td>2.58</td>
</tr>
<tr>
<td>Thompson Cr.</td>
<td>0.05</td>
<td>-0.82</td>
<td>0.994</td>
<td>5.68</td>
<td>0.03</td>
<td>-0.65</td>
<td>0.959</td>
<td>3.23</td>
</tr>
<tr>
<td>Susitna R. nr.Talk.</td>
<td>0.02</td>
<td>-1.03</td>
<td>0.897</td>
<td>2.75</td>
<td>0.01</td>
<td>-1.03</td>
<td>0.878</td>
<td>2.49</td>
</tr>
</tbody>
</table>
3.1.2 $\tau^*_{rD50tot}$ - Derived from total bedload transport rates

The reference shear stress values $\tau^*_{rD50tot}$ derived from total bedload transport rates $> 4$ mm (see section (2.3.2) and Table (2)) almost exactly correspond to the $\tau^*_{rD50}$ derived from fractional transport rates for the Main Dataset (Figure (5)), but deviate somewhat for the HS Dataset. The general trend of $\tau^*_{rD50}$ with channel slope for the Main Dataset is not strongly affected by whether $\tau^*_{rD50}$ was derived from fractional or total transport rates (Equations (10, 11)).

$$\tau^*_{rD50tot} = 0.495S^{0.47} \text{ for } S > 0.01; \; r^2 = 0.67$$  \hspace{1cm} (11)
Figure 3: Dimensionless reference shear stress ($\tau^*_{rD50}$) related to stream characteristics. $\tau^*_{rD50}$ derived from total acting bed shear stress $\tau^*_{D50}$ and fractional transport rates (see Section 2.3.2); $\tau^*_{rD50}$ corresponds to the coefficient given in the hiding function of Figure 2b. a) channel bed slope, with the empirical relations of Bunte et al., [2013], Lamb et al., 2008; and Mueller et al., [2005]; b) Sand content with the empirical relation of Wilcock and Crowe [2003]; c) $D_{84Surf}$ of the surface grain size distribution (GSD); d) $D_{50Surf}/D_{50Sub}$ as a measure of the degree of armouring with the empirical relation of Efthymiou [2012]; e) $D_{84Surf}/D_{30Surf}$ as a measure of the width of the bed surface GSD; f) Geometric standard deviation of the bed surface GSD with the empirical relation of Gaeuman et al., [2009].
Figure 4: Dimensionless reference shear stress ($\tau_{*rD50}^*$) related to stream characteristics. $\tau_{*rD50}^*$ derived from the reduced effective bed shear stress $\tau_{*rD50}$ ($RR2011$) and fractional transport rates. a) channel bed slope, with the empirical relations of Bunte et al., [2013] (dotted line), Lamb et al., [2008] (gray line); and Mueller et al., [2005] (thin black line); b) Sand content with the empirical relation of Wilcock and Crowe [2003] (WC2003); c) $D_{84Surf}$ of surface grain size distribution (GSD); d) $D_{50Surf}/D_{50Sub}$ as a measure of the degree of armouring with the empirical relation of Efthymiou [2012]; e) $D_{84Surf}/D_{30Surf}$ as a measure of the width of the bed surface GSD; f) Geometric standard deviation of the bed surface GSD with the empirical relation of Gaeuman et al., [2009].
Figure 5: Dimensionless reference shear stress ($\tau^*_{rD50}$) related to channel bed slope for a) the total acting bed shear stress $\tau^*_{rD50}$ and b) the reduced effective shear stress $\tau^{*'}_{rD50}$ (RR2011). Filled circles correspond to $\tau^*_{rD50}$ values from analyzing fractional transport rates (see reference approach Section (2.3.2) and Figure (2b)). Crosses indicate $\tau^*_{rD50}$ derived from analysis of total dimensionless transport rates. The black line was fitted to the Main Dataset, and the gray line to the HS Dataset. The dashed blue line in b) corresponds to the median $\tau^*_{rD50}$ from Main Dataset.

Figure 6: Boxplots of hiding exponent $b$, derived using total boundary shear stress $\tau^*$ and the reduced effective stress $\tau^{*'}$ (RR2011). Dark blue boxes are based on the Main Dataset values including the bedload trap, vortex and moving bedload basket data. Light gray boxes represent $b$ values of the Helley and Smith data. Boxplot edges represent the 25th and 75th percentiles, whiskers include data within ±2.7 standard deviations, and outliers are presented as red crosses. The gray line represents the hiding exponent $b$ as given by Wilcock and Crowe [2003] for $D<D_{50Surf}$. 
Mixed-size sediments, macro-roughness and bedload transport

Figure 7: Hiding exponent $|b|$ related to the $D_{50\text{Surf}}/D_{50\text{Sub}}$ ratio (degree of armouring) for a) the total boundary shear stress $\tau^*$ and b) the reduced effective shear stress $\tau^{**}$ according to RR2011. Equal mobility is implied by $|b|=1$; full size selectivity is implied by $|b|=0$.

A correlation of the hiding exponent $|b|$ with the $D_{84}/D_{16}$ ratio [Bathurst, 1987] or the flow regime (rainfall or nival) [Bathurst, 2013] could not be identified for the streams of the Main Dataset (all streams are dominated by snowmelt/glacier melt runoff, except the Erlenbach).

Because the available bedload transport data pertain mainly to low and moderate flow conditions, almost no fractional bedload transport rates were measured for grain sizes larger than the $D_{50\text{Surf}}$. This prevents a comparison with two-trend formulas of the hiding exponent $b$ towards the $D_i/D_{50\text{Surf}}$ ratio as derived from flume experiments by Wilcock and Crowe [2003] (Appendix, Equation (A4), or see a similar approach for the Trinity River by Gaeuman [2009]).

3.3. Bedload transport rating curves

3.3.1 Fractional bedload transport

The reference shear stress per grain size fraction $\tau^{*}_{ri}$ derived in step (ii) in section (2.3.2) was used to collapse fractional bedload transport rates computed from the total boundary shear stress approach and the effective shear stress approach according to RR2011 (Equation (6)). Power law exponents $m_i$ of collapsed fractional transport relations ($W^*=0.002\tau/\tau^{*}_{ri}$) are highly variable for all streams (Figure (8)). Whereas the median values of $m_i$ of the size classes larger than 4 mm plot more or less on the same level, the median $m_i$ of the 4 mm size class is slightly larger. In further analysis, we ignored the variability of $m_i$ with $D$ for the smallest size class and we determined the median exponent $m_{i50}$ of the fractional exponents $m_i$ for each individual stream (See Table (2) for total shear stress and effective shear stress RR2011, and Table (S1) for effective shear stress according to WC2009). Whereas the non-dimensional rating curves of the total boundary stress approach are significantly steeper (median exponent $m_{i50}=13.7$, Main Dataset only) than the exponent of 7.5 given in the Wilcock and Crowe [2003] equation, the rating curve exponents of the effective shear stress are on average (median $m_{i50}=7.3$, Main Dataset only) almost similar to the Wilcock and Crowe [2003] exponent (Figure (8, 9)).
Chapter III

Figure 8: Boxplots of power law exponents $m_i(W^\ast_i=0.002\tau^\ast_i/\tau^\ast_{ri}m_i)$ averaged over the study streams related to grain-size class, derived from a) the total shear stress and b) the effective shear stress according to Rickenmann and Recking [2011]. The blue thick lines represent the median exponent $m_{i50}$ averaged over all grain size classes and streams of the Main Dataset. The dashed lines represent the exponent given by Wilcock and Crowe [2003]. Boxplot edges represent the 25th and 75th percentiles, whiskers include data within ±2.7 standard deviations, and outliers are presented as red crosses.

Figure 9: Boxplots of exponents $m_{i50}$ (median of fractional exponents $m_i$) and $m_{tot}$ of power laws fitted to collapsed dimensionless bedload transport rates ($W^\ast_i=0.002\tau^\ast_i/\tau^\ast_{ri}$ and $W^\ast_{tot}=0.002\tau^\ast_{D58Surf}/\tau^\ast_{D58Surf}$, respectively). The two left Boxes are based on the total acting bed shear stress $\tau^\ast$, the two right boxes are based on the reduced effective acting bed shear stress of $\tau^\ast_{eff}$ (RR2011). Dark blue boxes are based on the bedload trap, vortex and moving bedload basket data. Light gray boxes represent $m$ values of the Helley and Smith data. Boxplot edges represent the 25th and 75th percentiles, whiskers include data within ±2.7 standard deviations, and outliers are presented as red crosses.
Figure 10: Similarity collapse of fractional transport rates based on $\tau^*_{ri}$ (derived from the reference approach of section (2.3.2), step (ii)). a) Collapse based on the total boundary shear stress $\tau^*$. b) Collapse based on the reduced effective bed shear stress $\tau^*'$ according to Equation 6 (RR2011) [Rickenmann and Recking, 2011].

Furthermore, the rating curve exponents $m_i$ are generally higher for the Main Dataset than for the HS Dataset, which is valid both for the total boundary shear stress and also the effective shear stress approach (Figure (8, 9)). The large variability of $m_{i50}$ could not be explained by any of the studied stream characteristics, except for weak correlations with the catchment area $A$ ($m_{i50}=15.6*A^{0.15}$; $r^2=0.29$) and bankfull flow $Q_{bklf}$ ($m_{i50}=26.3*Q_{bklf}^{-0.49}$; $r^2=0.35$) that emerge only for the complete dataset (within the Main Dataset only, no correlations could be identified). Figure (10) shows the collapsed fractional transport rates based on the reference shear stress per grain size fraction $\tau^*_{ri}$ for the total shear stress and the effective shear stress (RR2011) case (See Figure (S6) for WC2009). It is evident that the increase of dimensionless fractional transport rates $W^*_i$ is significantly steeper for the total shear stress case compared to the effective shear stress case. For high shear stresses ($\tau^*/\tau^*_{ri} \sim 1.3$) the collapsed dimensionless transport rates, especially for the HS data points, approximately follow the asymptotically approach of the Wilcock and Crowe [2003] ($WC2003$) equation towards a constant mean value, for both the total- and effective-shear-stress approaches (Figure (10)). However, dimensionless transport rates for $\tau^*/\tau^*_{ri} \sim 1.3$ tend to plot up to two orders of magnitude below the $WC2003$ equation, and up to two orders of magnitude above the $WC2003$ equation for smaller ratios of $\tau^*/\tau^*_{ri}$. Generally, the scatter of the collapsed fractional transport rates is larger for the effective shear stress approach compared to the total shear stress approach.
3.3.2 Total bedload transport rates

Non-dimensional rating curves were also fitted for the total dimensionless bedload transport rates $W_{tot}^*$ (>4mm) vs. $\tau_{D50}^* \tau_{D50}$ for both the total- and effective-shear-stress (RR2011) approaches (Table (3) and Table (S1), see also Figures (S7) and (S8)). The power-law exponents $m_{tot}$ for total transport rates are 14% – 47% higher than the exponents $m_{i50}$ fitted to the collapsed fractional rating curves (Figure (9)). Otherwise, the behavior of $m_{tot}$ is the same as for $m_i$: the total boundary shear stress derived $m_{tot}$ values are higher than the effective shear stress derived values, and the Main Dataset $m_{tot}$ values are generally higher than the HS Dataset values (Figure (9)).

3.4. Implication for bedload transport prediction

Our findings concerning the reference shear stress, the relative size effects and steepness of the bedload transport rating curves, as well as the effects of using total boundary shear stress or an effective shear stress, led us to set up dimensionless bedload transport models, similar to the original form of the Wilcock and Crowe [2003] (WC2003) equation. Rather than proposing an improved bedload transport equation for steep mountain streams, our main aim is to demonstrate how the effective shear stress estimate and the observed steep increase of bedload transport rates with increasing flow energy affect bedload transport prediction. The following six approaches were used for bedload transport prediction:

(a) The original WC2003 equation (Equations (A1-A4)) in combination with the total boundary shear stress (Figure (11a))

(b) The original WC2003 equation (Equations (A1-A4)) in combination with the effective shear stress RR2011 (Figure (11b))

(c) A modified WC2003 equation in combination with the total boundary shear stress, $\tau_{rD50}$ from Equation (10), $\tau_{ri}$ from Equation (A4) and a median exponent $m_{i50}$=13.7 for the Main Dataset for $\tau_{r}/\tau_{ri} < 1.33$ (Equation (12), Figure (11c))

$$W_i^* = \begin{cases} 0.002 \tau_{r}/\tau_{ri}^{1.37} & \text{for } \tau_{r}/\tau_{ri} < 1.2 \\ 14\left(1 - \frac{0.85}{\tau_{r}/\tau_{ri}}\right)^{4.5} & \text{for } \tau_{r}/\tau_{ri} \geq 1.2 \end{cases}$$ (12a,b)

(d) A modified WC2003 equation in combination with the effective shear stress RR2011, a constant reference shear stress of $\tau_{rD50} = 0.04$, $\tau_{ri}$ from Equation (A4), and a median exponent $m_{i50}$=7.3 for the Main Dataset for $\tau_{r}/\tau_{ri} < 1.33$ (Equation (13), Figure (11d))

$$W_i^* = \begin{cases} 0.002 \tau_{r}/\tau_{ri}^{7.3} & \text{for } \tau_{r}/\tau_{ri} < 1.33 \\ 14\left(1 - \frac{0.894}{\tau_{r}/\tau_{ri}}\right)^{4.5} & \text{for } \tau_{r}/\tau_{ri} \geq 1.33 \end{cases}$$ (13a,b)
(e) a modified WC2003 calculating total transport rates equation in combination with the total boundary shear stress \(\tau^*_{D50}\) from Equation (11) and a median exponent \(m_{tot} = 16.1\) for the Main Dataset for \(\tau^*/\tau^*_{ri} < 1.33\) (Equation (14), Figure (11e))

\[
W^*_{tot} = \begin{cases} 
0.002 \frac{\tau^*_{D50}}{\tau^*_{D50}}^{16.1} & \text{for } \tau^*/\tau^*_{ri} < 1.33 \text{ and } D=4\text{mm} \\
14 \left(1 - \frac{0.894}{\tau^*_{D50}/\tau^*_{D50}}\right)^{4.5} & \text{for } \tau^*/\tau^*_{ri} \geq 1.33 \text{ and } D=4\text{mm}
\end{cases}
\]  

(14a,b)

(f) a modified WC2003 calculating total transport rates equation in combination with the effective shear stress RR2011, a constant reference shear stress of \(\tau^*_{D50} = 0.03\), and a median exponent \(m_{tot} = 8.4\) for the Main-Dataset for \(\tau^*/\tau^*_{ri} < 1.33\) (Equation 15, Figure (11f))

\[
W^*_{tot} = \begin{cases} 
0.002 \frac{\tau^*_{D50}}{\tau^*_{D50}}^{8.4} & \text{for } \tau^*/\tau^*_{ri} < 1.33 \text{ and } D=4\text{mm} \\
14 \left(1 - \frac{0.894}{\tau^*_{D50}/\tau^*_{D50}}\right)^{4.5} & \text{for } \tau^*/\tau^*_{ri} \geq 1.33 \text{ and } D=4\text{mm}
\end{cases}
\]  

(15a,b)

Fractional transport rates >4 mm for approaches (a-d) were summed to total transport rates in Figure (11a-11d) to simplify the presentation. Note that for the Riedbach and Erlenbach only fractional transport rates were summed up with size fractions larger than the minimum sampled grain sizes of 6 mm and 10 mm, respectively. The expected error resulting from omitting fine gravel in those two cases decreases with increasing sampled transport rates and might amount to a factor of 2-5 for the smallest transport rate and a factor of < 0.2 for the largest. These errors are negligible in the context of the following analysis. For all fractional transport calculations, the hiding exponent of Wilcock and Crowe [2003] (Equation (A4)) was used. The rationale for using modified Wilcock and Crowe [2003] equations with two trends for \(\tau^*/\tau^*_{ri} < 1.33\) and \(\tau^*/\tau^*_{ri} \geq 1.33\) was based on the observation that dimensionless transport rates appear to asymptotically approach a constant value for high \(\tau^*/\tau^*_{ri}\). However we did not fit new equations for \(\tau^*/\tau^*_{ri} \geq 1.33\), due to the large scatter of the collapsed transport rates for \(\tau^*/\tau^*_{ri} \geq 1.33\) derived from the field data. Therefore, we kept the definition that \(W^*_i\) asymptotically approaches the value of 14 as defined from flume data by Wilcock and Crowe [2003], and we slightly modified the parameters and changed the \(\tau^*/\tau^*_{ri}\) criterion to 1.33 to yield an appropriately shaped transport function (red lines, Figure (10)).

The predictive accuracy was defined by a score representing the percentage of data that lay within one order of magnitude of the measured bedload transport rate \((SC_{10}, 0.1 < q_{calc}/q_{meas} < 10)\) and a second score representing the percentage within three orders of magnitude of the measured rate \((SC_{1000}, 0.001 < q_{calc}/q_{meas} < 1000)\), where \(q_{calc}\) is the computed bedload transport rate and \(q_{meas}\) is the measured unit bedload transport rate. The predictive accuracy is lowest for the WC2003 approach coupled with an unredused boundary shear stress (Figure (11a)), and inaccuracy increases with channel bed gradient (Figure (12a)). Using the WC2003 equation in combination with an effective shear stress [Rickenmann and Recking, 2011; Wilcock et al., 2009] significantly improves bedload transport predictions (Figures (11b) and (S9a)) and lessens slope-related overestimations (Figures (12b) and (S10a)), although predictions for the Main Dataset streams in particular remain overestimated (Figure (11b)).
Table 3: Dimensionless fractional bedload rating curves for total bedload transport rates > 4mm derived from total boundary shear stress $\tau^*_{D50}$ and the reduced effective shear stress $\tau^*_{D50}'$ according to Rickenmann and Recking [2011] (RR2011).

<table>
<thead>
<tr>
<th>Stream</th>
<th>Total boundary shear stress $W^<em>_{tot} = 0.002(\tau^</em><em>{D50}/\tau^*</em>{D50}rD50tot})^{m_{tot}}$</th>
<th>Effective shear stress (RR2011) $W^<em>_{tot} = 0.002(\tau^</em><em>{D50}/\tau^*</em>{D50}rD50tot})^{m_{tot}}$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$\tau^*_{D50tot}$</td>
<td>$m_{tot}$</td>
</tr>
<tr>
<td>Erlenbach</td>
<td>0.18</td>
<td>7.66</td>
</tr>
<tr>
<td>Riedbach</td>
<td>0.11</td>
<td>15.96</td>
</tr>
<tr>
<td>St. Louis Cr.</td>
<td>0.08</td>
<td>13.38</td>
</tr>
<tr>
<td>Cherry Cr.</td>
<td>0.11</td>
<td>24.93</td>
</tr>
<tr>
<td>Ltl.Granite Cr.A</td>
<td>0.06</td>
<td>32.78</td>
</tr>
<tr>
<td>E.St.Louis Cr.A</td>
<td>0.13</td>
<td>22.61</td>
</tr>
<tr>
<td>E.St.Louis Cr.B</td>
<td>0.14</td>
<td>22.62</td>
</tr>
<tr>
<td>Halfmoon Cr.</td>
<td>0.08</td>
<td>16.31</td>
</tr>
<tr>
<td>Hayden Cr.</td>
<td>0.11</td>
<td>18.33</td>
</tr>
<tr>
<td>E.Dallas Cr.</td>
<td>0.06</td>
<td>12.69</td>
</tr>
<tr>
<td>Fool Cr.</td>
<td>0.08</td>
<td>27.44</td>
</tr>
<tr>
<td>NF Swan Cr.</td>
<td>0.16</td>
<td>9.31</td>
</tr>
<tr>
<td>Ltl.Granite Cr.B</td>
<td>0.05</td>
<td>15.79</td>
</tr>
<tr>
<td>Oak Cr.</td>
<td>0.04</td>
<td>10.26</td>
</tr>
<tr>
<td>Big W.R..nr.Ket.</td>
<td>0.04</td>
<td>8.58</td>
</tr>
<tr>
<td>Boise R.</td>
<td>0.05</td>
<td>5.15</td>
</tr>
<tr>
<td>Dollar Cr.</td>
<td>0.08</td>
<td>4.92</td>
</tr>
<tr>
<td>Herd Cr.</td>
<td>0.03</td>
<td>5.33</td>
</tr>
<tr>
<td>Lochsa R.</td>
<td>0.03</td>
<td>8.00</td>
</tr>
<tr>
<td>MF Red R.</td>
<td>0.05</td>
<td>2.30</td>
</tr>
<tr>
<td>MF Salmon R.</td>
<td>0.03</td>
<td>16.07</td>
</tr>
<tr>
<td>NFCl.Water R.</td>
<td>0.02</td>
<td>8.34</td>
</tr>
<tr>
<td>Rapid R.</td>
<td>0.05</td>
<td>5.49</td>
</tr>
<tr>
<td>Salmon Rbl.Ynk.F</td>
<td>0.04</td>
<td>12.05</td>
</tr>
<tr>
<td>Salmon R. nr. Obs.</td>
<td>0.04</td>
<td>10.58</td>
</tr>
<tr>
<td>Salmon R. nr. Shp.</td>
<td>0.02</td>
<td>10.09</td>
</tr>
<tr>
<td>Selway R.</td>
<td>0.02</td>
<td>11.61</td>
</tr>
<tr>
<td>SF Payette R.</td>
<td>0.04</td>
<td>3.85</td>
</tr>
<tr>
<td>Squaw Cr.USGS</td>
<td>0.05</td>
<td>7.51</td>
</tr>
<tr>
<td>Thompson Cr.</td>
<td>0.06</td>
<td>6.38</td>
</tr>
<tr>
<td>Susitna R. nr.Talk..</td>
<td>0.03</td>
<td>6.46</td>
</tr>
<tr>
<td>Talkeetna Rnr.Talk..</td>
<td>0.01</td>
<td>5.15</td>
</tr>
<tr>
<td>Lo. NF Cabin Cr.</td>
<td>0.26</td>
<td>3.03</td>
</tr>
<tr>
<td>Buffalo Cr.</td>
<td>0.01</td>
<td>5.57</td>
</tr>
<tr>
<td>Horse Cr.</td>
<td>0.01</td>
<td>4.42</td>
</tr>
<tr>
<td>Lo. SF Williams F.</td>
<td>0.07</td>
<td>5.80</td>
</tr>
<tr>
<td>Williams F.</td>
<td>0.02</td>
<td>9.40</td>
</tr>
<tr>
<td>Median Main-Data</td>
<td>0.09</td>
<td>16.14</td>
</tr>
<tr>
<td>Median HS-Data</td>
<td>0.04</td>
<td>6.38</td>
</tr>
</tbody>
</table>
Mixed-size sediments, macro-roughness and bedload transport

Figure 11: Bedload transport predictions (> 4 mm) based on the total acting bed shear stress $\tau^*$ (left) and the reduced effective acting bed shear stress $\tau''$ (RR2011, right) for a & b) the Wilcock and Crowe [2003] equation; c) Equation (12) in combination with a reference shear stress based on channel bed slope (Equation (10)) derived from accumulated fractional calculations; d) Equation (13) in combination with a constant reference shear stress $\tau_{rD50}^* = 0.04$ derived from accumulated fractional calculations; e) Equation (14) in combination with a reference shear stress based on channel bed slope (Equation (11)) derived from total calculations; f) Equation (15) in combination with a constant reference shear stress $\tau_{rD50}^* = 0.03$ derived from total calculations.
Chapter III

Figure 12: Discrepancy ratios (calculated to measured unit transport rates) related to channel slope. Calculations based on the total acting bed shear stress $\tau^*$ (left) and the reduced effective acting bed shear stress $\tau'^*$ ($RR2011$, right) for a & b) the Wilcock and Crowe [2003] equation; c) Equation (12) in combination with a reference shear stress based on channel bed slope (Equation (10)) derived from accumulated fractional calculations; d) Equation (13) in combination with a constant reference shear stress $\tau^{*,rD_{50}}=0.04$ derived from accumulated fractional calculations; e) Equation (14) in combination with a reference shear stress based on channel bed slope (Equation (11)) derived from total calculations; f) Equation (15) in combination with a constant reference shear stress $\tau^{*,rD_{50}}=0.03$ derived from total calculations.
Mixed-size sediments, macro-roughness and bedload transport

For all six models tested in Figure (11), the predictions derived from models calibrated to the field data (Main Dataset) still display large scatter. There is a weak tendency for bedload transport predictions that include corrections for macro roughness (Figure (11d) and (11f)) to perform slightly better compared to the total boundary shear stress and the variable shear stress (Equation (10)) predictions. No substantial differences could be observed between transport rates predicted from total and fractional transport calculations.

4. Discussion

4.1. Reference shear stress $\tau^*_{rD50}$

It is evident from previous studies that the concepts of the critical and the reference shear stress, describing the initiation of motion, are closely related to each other [Bunte et al., 2013; Mueller et al., 2005; Wilcock et al., 2009]. This finding is supported by our analysis, since the reference shear stress relation plots very close to the critical shear stress relation derived by Bunte et al. [2013] from largely the same data (Figure (3a)). It could also be shown that the relationship between $\tau^*_{rD50}$ and channel slope occupies a range similar to other reference- and critical-stress relations (Figure 3a) [Mueller et al., 2005; Lamb et al., 2008]. However, we only found positive correlations between $\tau^*_{rD50}$ and channel slope if $\tau^*_{rD50}$ was derived from total boundary shear stress, as it was the case for values from other studies presented here. If an effective shear stress is used [Rickenmann and Recking, 2011; Wilcock et al., 2009] this positive trend disappears. Reference shear stress values approach a median of $\tau^*_{rD50} = 0.03$ for the Main Dataset (Table (3) and (S1)), and the values vary in a similar range as for less steep channels, where effective shear stresses are much closer to total shear stresses.

The finding that the reference shear stress remains independent of channel slope when derived from a reduced effective shear stress is in contrast to the results by Prancevic et al. [2014] from flume experiments on critical Shields values at steep slopes. They showed that critical Shields values (derived from total boundary shear stress, Equation (1)) increased with increasing bed slope, in agreement with the relation of Lamb et al. [2008]. Prancevic et al. [2014] assumed that energy losses due to form roughness can be neglected in their experiments. However, considering that in their flume experiments, the relative flow depth $h_r$ ($h_r = h/D$ where $h$ is flow depth and grain size $D=1.5$ cm) was typically between 0.13-1.7 but mostly <1.01, and that there was some subsurface flow through the pore space of the relatively coarse grains, we suggest that their critical Shields values are also affected by energy losses from macro-roughness or form-roughness.

The $D_{84Surf}/D_{30Surf}$ dependence obtained from the total boundary shear stress approach (Figure (3e)) is probably also affected by correlations between $D_{84Surf}/D_{30Surf}$ and channel slope, because grain size distributions typically widen with increasing channel slope [e.g., Gomez et al., 2001; Knighton, 1980; Rice and Church, 1998]. But despite this correlation, our analysis suggests that the $D_{84Surf}/D_{30Surf}$ ratio also directly affects sediment mobility, which is consistent with literature results [Ferguson, 2012].

4.2. Hiding exponent $b$

The range of $b$ values found in this study (Figure (6)) is similar to the range of $b$ values of -0.6 to -1 reported in the literature [e.g., Andrews, 1994; Ashworth and Ferguson, 1989;
Chapter III

Bathurst, 2013; Bunte et al.; 2013; Ferguson, 1994; Gaeuman et al., 2009; Mao et al., 2008; Parker, 1990; 2008; Recking, 2009]. Median $b$ values close to -0.92 to -0.96 indicate a tendency towards equal mobility. With respect to the reference shear stress $\tau_{rD50}$, the hiding exponent $b$ is not considerably affected when derived from the total boundary or effective shear stress approach.

The hiding exponent $b$ plays an important role for the calculation of the transport rates of individual particle size fractions. However, to date, no strong predictive parameter could be identified which describes the hiding and exposure effects. Some studies suggested that the hiding exponent could be related to bed, flow or bedload transport characteristics such as particle sphericity [Carling, 1983], the $D_{84}/D_{16}$ ratio [Bathurst, 1987], the flow regime controlling bed consolidation [Bathurst et al., 2013] or the steepness of the critical flow curve $Q_c = f(D_{max})$ [Bunte et al., 2013]. However, none of those dependencies could be confirmed by independent datasets yet. Also, within this study we could not demonstrate that one of these previously introduced stream bed, flow or bedload transport parameters or other predictive parameters could explain the variation of the hiding parameter $b$ (Figures (S3, S4 and S5)). At least, we identified an inverse relation of the hiding exponent $|b|$ with the degree of armoring $D_{50Surf}/D_{50Sub}$ ($r^2 = 0.44$ and 0.48, respectively) (Figure (7)), meaning that the larger the $D_{50Surf}/D_{50Sub}$ ratio, the closer the exponent is to the size-selective case. This observation is plausible, because if finer particles are abundant and more mobile compared to coarse ones then the channel bed should develop a coarse armor layer.

4.3. Dimensionless bedload transport rating curves

Why rating curve exponents $m_i$ (and $m_{i50}$ respectively) and $m_{tot}$ were smaller when based on effective rather than total boundary shear stress (Figures (9) and (10)) might not be intuitively apparent. This is because the range of $\tau^*/\tau^*_r$ (both for fractional and total transport rates) is larger for a given stream and a given flow discharge range when using a constant reference shear stress as compared to an effective shear stress approach. Using the Equations (6) or (7), the effective shear stress is reduced as a function of relative flow depth (total boundary shear stress is reduced more strongly at shallower flows compared to deeper flows). By contrast, for the total boundary shear stress approach, the available flow energy depends on the absolute, but not on the relative flow depth.

The larger $m_i$ and $m_{tot}$ exponents for the Main Dataset streams compared to the HS Dataset streams might be partly explained by uncertainties associated with the Helley-Smith (HS) bedload samples. Bunte et al. [2008] found that bedload transport rating curves determined with bedload traps are generally steeper than the rating curves determined from HS samplers. Whereas bedload transport rates measured with bedload traps increase steeply with discharge, bedload transport rates measured with HS samplers at the same sites yielded higher transport rates during low-flow conditions and increased less steeply with discharge. The oversampling by the HS sampler at low flows has been attributed to short sampling times, to the flared HS sampler design, and to involuntary particle pick-up during sampler placement [Bunte and Abt, 2005].

Apart from the differences in rating curve exponents $m$ ($m_i, m_{i50}$ or $m_{tot}$, respectively), the variability of $m$ within each group (total boundary shear stress vs. reduced shear stress; Main Dataset vs. HS Dataset; fractional vs. total transport rate calculations) is large as well.
Mixed-size sediments, macro-roughness and bedload transport

However, we could not identify a strong relationship between \( m \) and any of the grain size, streambed or flow characteristics (in contrast to, e.g. the mean shear stress of a study site as reported by Diplas and Shaheen [2007]) that could explain these variations. Indeed, we found weak negative correlations of \( m \) with catchment area and bankfull flow (Section 3.3.1), but only for the complete dataset, and this result might be different when considering that HS bedload transport rating curves (which represent mainly the larger catchments in this study) have generally smaller rating curve exponents. Furthermore, in the literature we found no suggestion that \( m \) might depend on stream or hydraulic characteristics, or on whether the \( m \) exponents are derived from flumes [e.g., Proffitt and Sutherland, 1983; Recking, 2010; Wilcock and Crowe, 2003] or field data [e.g., Recking, 2010; Wilcock, 2001], with \( m \) values ranging in general from about 5 to 14.

The observation in this study that dimensionless transport rates \( W^*_i \) approach a constant value for high \( \tau^*_i/\tau^*_r \) values (Figure (10)) corresponds to previous observations. Several studies, mainly using flume data [Diplas and Shaheen, 2007; Meyer-Peter and Mueller, 1948; Parker, 1990; Rickenmann, 1991; Wilcock and Crowe, 2003], have shown that transport observations asymptotically approach a relation \( q_{b*} \sim (\tau^*)^{1.5} \) for high shear stresses \( \tau^*/\tau^*_r \) with a pre-factor (coefficient) in the range of about 4 to 14 [Wilcock and Crowe, 2003; Wong and Parker, 2006; Wilson, 1987].

4.4. Bedload transport prediction

Among the six versions of the Wilcock and Crowe [2003] transport model compared in this study, the original approach (WC2003) that employs an unreduced boundary shear stress (Figure (11a)) had the lowest predictive accuracy. This is not surprising, because the equation was not developed for steep streams; it does not take potential energy losses into account and thus tends to overestimate bedload transport rates, especially at steep slopes (Figure (12a)). Using the WC2003 equation in combination with an effective shear stress as by Rickenmann and Recking [2011] (RR2011) significantly improves bedload transport predictions and lessens overestimations on steep slopes (Figures (11b) and (12b)). However, especially in the Main Dataset, transport rates tend to be overestimated, which corresponds to the observation from Figure (4b) that the reference shear stress based on the sand content underestimates the reference shear stress and overestimates the mobility of gravel particles.

The predictions of the modified WC2003 equations shown in Section (3.4) (Equations (12 to 15), Figures (11c to 11f)) are still greatly scattered, however, using a reduced effective shear stress and a constant reference shear stress (Equations (13, 15), Figures (11d, 11f)) result in slightly better bedload transport predictions than using the total boundary shear stress and the variable reference shear stress from (Equations (12, 14)). Transport rates that were predicted based on total or fractional transport calculations did not substantially differ.

Despite the slight improvement of bedload transport predictions using one of the suggested methods (Figures (11c to 11f)), the uncertainties in predicted transport rates remain huge (up to roughly 3 orders of magnitude), especially when considering that these equations have been calibrated to the field data. These uncertainties indicate either that the measurement data are noisy or that all bedload transport equations are generally missing one or several very important parameters. Variability in measured transport rates is likely due to the typically complex field conditions and the difficulties of measuring transport rates and flow conditions.
that vary greatly in both time and space. However, we assume that the bedload trap, moving basket and vortex trap measurements yield some of the most accurate field data currently available for steep streams. It is also likely that there is still something missing in our transport calculations. For example, we do not consider sediment supply issues [e.g., Bathurst, 2007; Recking, 2012; Yager et al., 2012b] or different phases of sediment transport [e.g. Diplas and Shaheen, 2007; Recking, 2010] which are important factors controlling bedload transport rates in steep mountain streams and might explain some of the remaining uncertainties.

As mentioned above, the predictive accuracies of the fractional and the total bedload transport equations (Figure (11c, 11d) and Figure (11e, 11f), respectively) are comparable, mainly because of the dominant influence of the reference shear stress compared to the grain size interactions. However, if measured fractional transport rates are available, the derivation of a fractional transport model has a significant advantage compared to a total transport equation. The derivation of a fractional model does not depend on the minimum grain size measured at a field site, which can vary significantly depending on the measurement technique (e.g. from 0.25 mm in the King dataset to 10 mm at the Erlenbach). The differences in the minimum grain sizes make it difficult to calibrate total bedload transport equations across different streams. By contrast, further application of fractional transport models is not restricted to a certain range of grain sizes.

5. Conclusions

Using a broad field dataset, we showed that transport equations such as the Wilcock and Crowe [2003] equation, which were developed on flume data for lower stream gradients, can also be applied to steep mountain streams if one accounts for flow resistance due to macro-roughness and/or the increased demand of energy for initiation of motion.

From our field data it remains unclear how much of the flow energy available for bedload transport is lost due to flow resistance of macro-roughness elements in steep mountain streams and how much more energy is needed for particle entrainment of particles experiencing structural stability. Fortunately, for bedload prediction itself it does not matter whether one reduces the flow energy available using a flow resistance equation or whether one increases the critical threshold for particle entrainment (see Figures (11c to 11f)). Both approaches correct the available energy in the same direction (see Figure (1)) and avoid systematic overestimation of sediment transport rates typical for steep mountain streams [e.g., Bathurst et al., 1987; Chiari and Rickenmann, 2011; Lenzi et al., 1999; Rickenmann, 2001; Rickemann, 2012; Yager et al., 2007; Yager et al., 2012a]. However, the choice of the one or the other approach has further implications for bedload transport predictions, as the stress-dependent increase in bedload transport rates defined by the power law exponents \( m \) strongly depends on which approach was used. It is interesting that the power law exponent of 7.3 derived from using the effective shear stress according to Rickenmann and Recking [2011], and of 7.0 for the effective shear stress approach of Wilcock et al. [2009], are, on average for the Main Dataset streams, very close the exponent of 7.5 given in the Wilcock and Crowe [2003] equation. Based on this similarity, the Wilcock and Crowe [2003] equation may be used, in combination with an effective shear stress approach, in steep mountain streams.
Mixed-size sediments, macro-roughness and bedload transport

without considerable changes to the bedload transport relation, as would be necessary when using a reference shear stress that depends on channel slope.

Finally, bedload transport predictions do not significantly vary whether they are derived from calculating transport rates for individual grain fractions, or from calculating total transport. It appears that in comparison to the variability of reference or critical shear stress and the effects of macro-roughness on transport efficiency (and their associated uncertainties), hiding and exposure effects are relatively insignificant. Uncertainties arising from the hiding exponent derived for the streams considered in this study very well are almost canceled out if one calculates total transport rates from accumulated fractional transport rates. If one is interested in the transport of an individual size fraction, the accurate determination of \( b \) becomes more important. However, despite the broad range of stream characteristics investigated in this study, we could not identify a strong predictive parameter for the hiding exponent \( b \) which controls the degree of size selective transport. At least one can roughly estimate that \( b \) tends to move towards greater size-selectivity with an increasing degree of bed armoring \( (D_{50Surf}/D_{50Sub}) \).


The Wilcock and Crowe [2003] equation is based on flume measurements that cover a wide range of flow, transport rates and bed surface sediments. The model is based on previous surface-based transport models [Parker, 1990; Proffitt and Sutherland, 1983], however, its hiding function incorporates a nonlinear effect of sand content on gravel transport rates [Wilcock et al., 2001]. The dimensionless fractional transport rate \( W_i^* \) is defined as a function of \( \theta_d = \tau/\tau_{ri} \).

\[
W_i^* = \begin{cases} 
0.002\theta_d^{7.5} & \text{for } \theta < 1.35 \\
0.894 & \text{for } \theta \geq 1.35 
\end{cases}
\]  

(A1)

The reference shear stress \( \tau_{ri} \) of each grain size fraction is determined using a dimensional form of a hiding function (equivalent to Equation (1)), with its input parameters \( \tau_{r50} = \tau_{rDm} \) and \( beta \) (note \( beta \) corresponds to 1-\( b \), in the dimensional form of the hiding function).

\[
\tau_{ri} = \tau_{rDm} \left( \frac{D_i}{D_m} \right)^{beta} 
\]  

(A2)

\[
\tau_{rD50}^* = (0.021 + 0.015\exp[-20F_i])(\rho_i - \rho)gD_{50Surf} 
\]  

(A3)

\[
beta = \frac{0.67}{1 + \exp\left(1.5 - \frac{D_i}{D_{50Surf}}\right)} 
\]  

(A4)

Where \( F_i \) is the proportion of sand in surface size distribution, \( D_i \) is the grain size of fraction \( i \), and \( D_{50Surf} \) is the \( D_{50} \) of the streambed surface.
Chapter III

ACKNOWLEDGEMENTS

This study was supported by the CCES project APUNCH of the ETH domain and partly by SNF grant 200021_124634/1 to DR. The long term field data collection with bedload traps in the US was supported by the Stream Systems Technology Center (now National Stream and Aquatic Ecology Center) of the USDA Forest Service, Rocky Mountain Research Station, Fort Collins, CO, USA. We thank K. Swingle, B. Schmid, N. Federspiel and K. Steiner for field assistance. Please contact the authors if you are interested in the Swiss bedload data (DR) or the US bedload trap data (KB).

REFERENCES


Ashida, K., and M. Michiue (1973), Study on bed load transport rate in open channel flows paper presented at International Symposium on River Mechanics IAHR, Bangkok, Thailand.


Mixed-size sediments, macro-roughness and bedload transport


Efthymiou, N. P. (2012), Transient Bedload Transport of Sediment Mixtures under...
Disequilibrium Conditions - An Experimental Study and the Development of a New Dynamic Hiding Function, PhD Dissertation, Technische Universität München, München.


Klingeman, P. C. (1979), Sediment transport research facilities, Oak Creek, Oregon Rep., *Water Resources Research Institute, Oregon State University, Corvallis*.


Mixed-size sediments, macro-roughness and bedload transport

Research-Earth Surface, 113(F2), Doi 10.1029/2007jf000831.


Parker, G., P. C. Klingeman, and D. G. Mclean (1982), Bedload and Size Distribution in
Chapter III


Shields, A. (1936), Anwendung der Ähnlichkeitsmechanik und der Turbulenzforschung auf die Geschiebebewegung, Triltsch & Huther.


Stabilisation with Placed Blocks, Hydraulics Section, Central Laboratories, Works Corporation.


CHAPTER IV

Scaling relationships between bedload volumes, transport distances and stream power in steep mountain channels

Johannes M. Schneider, Jens M. Turowski, Dieter Rickenmann, Ramon Hegglin, Sabrina Arrigo, Luca Mao, James W. Kirchner

ABSTRACT - Bedload transport during storm events is both an agent of geomorphic change and a significant natural hazard in mountain regions. Thus predicting bedload transport is a central challenge in fluvial geomorphology and natural hazard risk assessment. Bedload transport during storm events depends on the width and depth of bed scour, as well as the transport distances of individual sediment grains. We traced individual gravels in two steep mountain streams, the Erlenbach (Switzerland) and Rio Cordon (Italy), using magnetic and radio frequency identification (RFID) tags, and measured their bedload transport rates using calibrated geophone bedload sensors in the Erlenbach and a bedload trap in the Rio Cordon. Tracer transport distances and bedload volumes exhibited approximate power-law scaling with both the peak stream power and the cumulative stream energy of individual hydrologic events. Bedload volumes scaled much more steeply with peak stream power and cumulative stream energy than tracer transport distances did, and bedload volumes scaled as roughly the third power of transport distances. These observations imply that large bedload transport events become large primarily by scouring the bed deeper and wider, and only secondarily by transporting the mobilized sediment farther. Using the sediment continuity equation, we can estimate the mean effective thickness of the actively transported layer, averaged over the entire channel width and the duration of individual flow events. This active layer thickness also followed approximate power-law scaling with peak stream power and cumulative stream energy, and ranged up to 0.57 m in the Erlenbach, broadly consistent with independent measurements.

1. Introduction

Bedload transport during flood events in mountain channels is an important agent of geomorphic change [Schumm, 1977], and an important natural hazard, which can cause enormous damage in mountainous regions [e.g., Badoux et al., 2013]. Estimates of bedload transport typically rely on measurements that require substantial effort and specially designed samplers, possibly involving construction within the streambed (e.g., vortex samplers [Klingeman, 1979], Birkbeck samplers [Reid et al., 1980] or moving bedload baskets [Rickenmann et al., 2012]), or that can be used only in specific locations for short intervals of time (e.g., Helley-Smith samplers or bedload traps [Helley and Smith [1971]; Bunte et al., 2004]). In unmeasured stream reaches, estimates of transported bedload volumes rely on empirical equations, which are normally developed from limited laboratory and field data and have large predictive uncertainties, especially in mountain streams [e.g., Bathurst et al., 1987;
Another approach for estimating bedload transport is based on the notion that the bulk bedload is controlled by individually moving particles. Typically, the movements of individual particles consist of a series of steps and rest periods due to the turbulent flow conditions and irregular bedforms [e.g., Einstein, 1937; Hassan et al., 1991; Lajeunesse et al., 2010]. The sum of those individual particle step lengths during a transport event results in a total transport distance. Information on transport distances (or velocities), combined with information on entrainment rates or dimensions of a mobile layer depth, enable the estimation of bedload volumes (or rates) during a flood event (or observation period) [Hassan and Ergenzinger, 2005; Haschenburger and Church, 1998; Houbrechts et al., 2012; Laronne et al., 1992; Liébault and Laronne, 2008; Sear et al., 2000; Wilcock, 1997; Wong and Parker, 2006; Wong et al., 2007]. To determine total transport distances in natural channels, particles have been traced for individual transport events or longer observations periods using, e.g., color markings, magnetic tracers, or radio frequency identification (RFID) transponders [e.g., Bunte and Ergenzinger, 1989; Gintz et al., 1996; Haschenburger and Church, 1998; Lamarre and Roy, 2008; Lamarre et al., 2005; Liébault et al., 2012; Schmidt and Ergenzinger, 1992]. Particle tracking data have been used to constrain transport distance distributions and their dependence on channel morphology or flood magnitude (e.g., the Einstein-Sayre-Hubbel model, gamma or exponential distributions [Bradley and Tucker, 2012; Einstein, 1937; Gintz et al., 1996; Hassan et al., 1991; Hassan et al., 2013; Liébault et al., 2012; Sayre and Hubbell, 1964]). However, few studies exist for mountain streams where measured bedload volumes and transport distances of individual particles are available for the same transport events. Lenzi [2004] provided data for bedload volumes and transport distances for individual transport events for the Rio Cordon (Italy) while Liebault and Laronne [2008] measured the total bedload yield for an entire tracer transport distance campaign in the Escanovette torrent (France). Houbrechts et al. [2012] provided measurements of bedload transport and transport distances for several larger and lower-gradient rivers in the Ardennes (Belgium).

Understanding how bedload volumes and transport distances scale with one another, and with the size of a flood event (as characterized, for example, by the peak flow or the total storm runoff) is important for predicting large transport events and also for understanding how they affect the streambed. To characterize the impact of transport events on the streambed we use the concept of the active layer [Hirano, 1971; Parker, 1991; Parker et al., 2000], which can be defined as ‘the portion of the streambed that is mobilized during floods competent to transport sediment’ [Haschenburger and Church, 1998]. The depth of the active layer may be relevant to fields as diverse as ecology (e.g., excavation of spawning gravels), engineering (e.g., scour of engineered structures in streams), and geomorphology (e.g., as a control on sediment delivery during flood events).

In this study we compiled a distinctive dataset on measured transport distances and bedload volumes in two steep mountain streams, the Erlenbach (Switzerland) and Rio Cordon (Italy, Lenzi et al., [2004]) for flood events with a broad range of flow magnitudes. We examine how transport distances and bedload volumes scale with the stream power of the peak discharge and the stream energy of the entire flood event, and compare our findings with literature results. Based on the transport distance and bedload data it was possible to back-calculate an active layer depth to estimate the effects of flow events on the streambed.
Chapter IV

2. Field sites and methods

2.1. Field sites

2.1.1. Erlenbach

The Erlenbach is a steep mountain stream (mean gradient: 0.17) draining an area of 0.7 km² in the Swiss pre-Alps (figure 1; table 1). The runoff regime is nivo-pluvial (snow and rain-dominated) with the largest transport events caused by extreme summer rainstorms. Occasional snowmelt and rain-on-snow events are of secondary importance for sediment transport.

In this study, transport events were observed covering a range of flow magnitudes with peak discharges ($Q_p$) of 0.5 – 10 m³/s. For later analysis we define transport events with $Q_p < 1.5$ m³/s, $1.5 \leq Q_p < 3$ m³/s, and $Q_p \geq 3$ m³/s as low-magnitude, moderate, and high-magnitude flood events, respectively. The corresponding recurrence intervals are approximately < 1 year, 1-3 years and > 3 years, respectively [Liechti, 2008].

<table>
<thead>
<tr>
<th>Table 1: Main characteristics of the Erlenbach and Rio Cordon catchments</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Erlenbach (CH)</strong></td>
</tr>
<tr>
<td>Basin elevation range [m]</td>
</tr>
<tr>
<td>Basin area [km²]</td>
</tr>
<tr>
<td>Mean slope/slope study reach [m/m]</td>
</tr>
<tr>
<td>Mean width [m]</td>
</tr>
<tr>
<td>Channel type</td>
</tr>
<tr>
<td>$D_{50surf}$/$D_{50surf}$/$D_{84surf}$ [mm]²</td>
</tr>
<tr>
<td>Lithology</td>
</tr>
<tr>
<td>Forested %</td>
</tr>
<tr>
<td>Discharge regime</td>
</tr>
<tr>
<td>RI2/RI5/RI10 [m³/s]²</td>
</tr>
<tr>
<td>Mean annual precipitation [mm]</td>
</tr>
<tr>
<td>Average bedload yield [m³ km⁻² y⁻¹]</td>
</tr>
</tbody>
</table>

² referring to the bed surface sediment

² Flood with recurrence interval (RI) of 2, 5 and 10 years [Liechti, 2008; Lenzi et al., 2004]
Flow discharge at the Erlenbach is measured directly upstream of a sediment retention basin, but because the main part of the Erlenbach tracer study reach was upstream of a small tributary (figure 1), we reduced the measured discharge proportionally to account for the reduced drainage area. The Erlenbach stream is characterized by a rough channel bed with a step-pool morphology and large relatively immobile boulders [Molnar et al., 2010]. The grain-size distribution (GSD) of the channel bed surface was determined by grid-by-number pebble counts in September 2012 using a variant of the Wolman [1954] procedure, according to Bunte and Abt [2001a, 2001b] and Bunte et al. [2009] (figure 2).

Bed material subsurface GSDs were determined by sieve analysis of three large volumetric samples taken using a three-sided plywood shield [Bunte and Abt, 2001b]. The GSDs of the transported bedload were determined by four sieved samples taken in the sediment retention basin from 1984 to 1987 [see also Rickenmann and McArdell, 2007]. The characteristic grain sizes $D_{25,\text{Surf}}/D_{50,\text{Surf}}/D_{84,\text{Surf}}$ of the streambed surface were 16/64/206 mm. In this paper, characteristic grain sizes of the streambed surface are subscripted with “Surf”, of the streambed subsurface with “Sub”, and of the bedload transported to the retention basin with “B”, respectively (e.g., $D_{50,\text{Surf}}$, $D_{50,\text{Sub}}$ and $D_{50,\text{B}}$).

2.1.2 Rio Cordon

Rio Cordon is a boulder-bed stream draining an area of 5 km$^2$ in the Italian Alps [see also Lenzi, 2004; Lenzi et al., 2006; Mao et al., 2010; Mao et al., 2009, and references therein], with a mean channel gradient of 0.13. Precipitation occurs mainly as snowfall from November to April and runoff is usually dominated by snowmelt in May and June, but summer and early autumn floods represent an important contribution to the flow regime. The peak discharges of the flood events analyzed during this study range from about 0.8 to 10 m$^3$/s.

The channel bed consists of step-pool, riffle-pool, and mixed reaches. The characteristic grain sizes of the streambed surface $D_{25,\text{Surf}}/D_{50,\text{Surf}}/D_{84,\text{Surf}}$ were identified by grid-by-number pebble counts as 38/90/262 mm, and the bed surface is strongly armored ($D_{50,\text{Surf}}/D_{50,\text{Sub}} \sim 3$) [Lenzi, 2004].
2.2. Characterization of hydraulic forcing

Bedload transport rates and distances are functions of hydraulic forcing, which in this study is characterized by stream power \[\textit{Bagnold}, 1966\]. Stream power has been shown to be more reliable than shear stress as a measure of hydraulic forcing \[\textit{e.g., Gomez and Church, 1989; Rickenmann, 2001}\], because estimating reach-averaged flow depth is difficult in steep mountain streams. Stream power per unit area \((\omega, \text{in W/m}^2)\) is derived from discharge \((Q)\) measurements and is defined as

\[
\omega = \frac{\rho g QS}{w} \tag{1}
\]

Where \(\rho\) is the fluid density \([\text{kg/m}^3]\), \(g\) is the acceleration due to gravity \([\text{m/s}^2]\), \(S\) is the channel slope \([\text{m/m}]\) and \(w\) is the flow width \([\text{m}]\).

Tracer transport distances, sediment volumes and the depth of the active layer are expressed in this study through functional relationships with the peak stream power (the unit excess stream power \(\omega_p-\omega_c\) in W/m\(^2\) \[see also \textit{Hassan and Church, 1992}\] at the peak discharge \(Q_p\)), and also with the cumulated stream power values per flood event, i.e., the sum of unit excess stream power values over the entire flood event \(\Sigma(\omega - \omega_c)\) which we term the "cumulative stream energy" in J/m\(^2\). Because the discharge measurements at the Erlenbach are conducted at one-minute time steps, we multiplied the one-minute values by 60 seconds for the cumulative values. The critical stream power \(\omega_c\) defining the initiation of motion is based on a critical discharge of 0.48 m\(^3\)/s for the Erlenbach \[\textit{Turowski et al., 2011}\] and a critical discharge of 0.65 m\(^3\)/s for Rio Cordon \[\textit{Lenzi, 2004}\].

Functional relationships between stream power, bedload transport, and tracer transport distances were derived using functional analysis on log-transformed values (for more details see \textit{Mark and Church}[1977]), because we expect errors in both the independent variable (stream power) and the dependent variables (bedload, transport distance, active layer depth). Simplifying the analysis, we have assumed that the errors in the x- and y-directions are equal. This is justified because we generally have strong correlations, and thus the apportionment of the errors between the x- and y-directions has only little effect on the outcome. In addition, the fitted coefficients were corrected for log-transformation bias \[\textit{Ferguson, 1986; Miller, 1984}\].

2.3. Estimation of mean transport distances

2.3.1 Erlenbach

2.3.1.1 Tracer particle seeding and recovery

In the Erlenbach two measurement campaigns with tracer particles were completed, a magnetic tracer campaign from 1994 to 1999 and a radio frequency identification (RFID) campaign in 2009 and 2010. The magnetic tracer technique has been applied for many years to track particle movements in gravel-bed streams \[\textit{e.g., Bunte and Ergenzinger, 1989; Gintz et al., 1996; Haschenburger and Church, 1998; Schmidt and Ergenzinger, 1992}\]. RFID transponders have recently enabled the tracking of individual stones using mobile antennas in wadeable gravel-bed streams \[\textit{e.g., Lamarre and Roy, 2008; Lamarre et al., 2005; Liébault et al., 2012}\].
The main advantage of using passive RFID transponders instead of magnetic tracers is that they are programmable with a unique identifier (ID) and can be identified down to a depth of 0.6 m without disturbing the streambed, regardless of whether there is water, rock, wood or mud in between [Schneider et al., 2010]. The term ‘tracer’ as used in this study refers to either magnetic or RFID-tagged particles.

During the magnetic tracer campaign, in June 1994 and May 1997, a total of 313 and 227 tracer particles, respectively, were placed onto the streambed 540 m upstream of the sediment retention basin (figure 1, location 11) [Schwer and Rickenmann, 1999; Schwer et al., 2010]. The tracers were located with a metal detector shortly after each sediment transport event whenever possible (survey periods 17 to 23, table 2).
The magnetic technique does not allow remote identification of individual particles. Buried tracers were uncovered, identified and replaced on the streambed surface above the location where they were found. In June 2009, 303 RFID tracers were placed onto the streambed 350 m upstream from the retention basin (figure 1, I2; figure 3). After the first transport event, 127 additional RFID tracers were placed at the same location. In May 2010, 142 RFID tracers were placed on the streambed 220m upstream from the retention basin (figure 1, I3), and 161 RFID tracers were placed 80m upstream from the retention basin (figure 1, I4; figure 3).

These shorter distances from the retention basin refer to a channel reach with less woody debris and therefore easier access. Thus it was hoped to increase the tracer recovery rates, which were generally low in the Erlenbach (see section 2.3.1.2). The positions of the tracer particles on (and in) the streambed were determined with a mobile RFID antenna, directly after flood events for which bedload transport was recorded by the geophone system (survey periods 1 to 16, table 2).

The GSD of the tracer particles corresponds to $D>D_{60B}$ of the sediment transported into the sediment retention basin, and to $D_{30Surf}<D<D_{75Surf}$ of the stream bed surface material (figure 2). A detailed description of the prepared particle size classes can be found in the supplementary material (section S1). Uncertainties introduced by the narrow GSD in determining the mean transport distances are discussed in section 2.3.1.3 and in the supplementary material (section S3). For further analysis we also included transport events in which the tracers were placed manually in their starting positions on the streambed; these include events 1, 2 and 8 of the RFID campaign and all events of the magnetic tracer campaign. We also neglected the fact that deeply buried particles might have been mobilized at a later time in a future flow event [Hassan et al., 2013]. For simplicity we assumed that the total transport distance of an individual grain results from multiple periods of transport and rest during a transport event, and therefore was not strongly affected by the arbitrary initial particle position, once a particle is entrained. In addition, we found no differences in transport distances for events in which the tracers started from seeded positions on the streambed surface or from naturally deposited positions (see also section 3.1.3, figure 6).
2.3.1.2 Tracer recovery rates

The recovery rates of tracer particles for the individual surveys were generally low in the Erlenbach, resulting in a limited number of data points (table 2) [see also Schneider et al., 2010]. These recovery rates (here called ‘total recovery rates’ and denoted as $P_t$), were calculated as the ratio of total particles found ($N_t$) to the total number of seeded particles, and averaged roughly 30% (min 1%, max 77%) across the 23 events that were analyzed.

This paper analyzes individual events, based on changes that occurred between successive pairs of surveys (typically performed shortly after each transport event). Therefore, we determined the number of particles located during each survey that were also located during the preceding survey ($N_{EvB}$) and a ‘event-based recovery rate’ ($P_{EvB}$), i.e. the recovery rate of particles found in two subsequent surveys ($P_{EvB0j} = 100 \cdot N_{EvB0j}/N_{t(i-1)}$). For the first survey after particle seeding, the event-based recovery rate corresponds to the total recovery rate. Before survey 2 (table 2), 127 additional particles were injected, thus the event based recovery rate of survey period two is based on the particles found in the previous event and the newly injected 127 particles.

To help in determining whether particles could not be found during a transport event because they were flushed out of the study reach or because they were buried at a depth outside the detection range of the antenna, we quantified the ‘re-emergence rate’ ($P_{RER}$) of the tracers. The re-emergence rate refers to tracer particles that could not be found after one transport event but were known to have moved in later events, because they were found in subsequent surveys or detected immediately upstream of the retention basin using a stationary RFID antenna (as developed by Schneider et al. [2010]). Due to technical problems, this antenna provided no data before September 2010; however, it did detect tracers as they were flushed out of the study reach in 2011 and 2012, so we could identify these as tracer particles that had remained in the study reach during 2009 and 2010.

Finally, the ‘event-based percentage’ of lost particles (those which either left the study reach or could not be detected anymore) was determined as $P_{Lost} = 100 - P_{EvB} - P_{RER}$. A detailed discussion of the low recovery rates can be found in the supplementary material, section S2.

As a consequence of the low recovery rates, for the further analysis of transport distances between successive particle surveys, we selected individual flood events for which (i) more than 10 tracer particles moved and (ii) less than ~30% of particles were lost.

2.3.1.3 Mean transport distances of tracer population and total bedload

We determined the mean transport distances of the tracer particles using the arithmetic mean of the individual tracer transport distances along the thalweg, considering only tracers that were known to have moved. The mean transport distance includes only tracers that moved, because later in the paper we compare this distance with the transported bedload volume, and that volume (by definition) includes only particles that moved in a given event. Using individual transport distances of tracers to estimate the mean transport distance of the bedload is problematic, because transport distance typically depends on particle size [cf. Church and Hassan, 1992; Haschenburger, 1996; Haschenburger and Church, 1998; Hassan et al., 1992; Lenzi, 2004; Wilcock, 1997] and the size distributions of the tracers and the
bedload differ in two respects: (1) the mean b-axis of the collected tracers may not always correspond to the mean b-axis of the tracer population, and (2) the tracer grain-size distribution (GSD), which ranges from 28 to 160 mm, is narrower than the average bedload GSD, which ranges from ~10 to 300 mm (figure 2). To correct for these sampling biases, i.e., to estimate transport distances of finer and coarser grains that were not represented by the tracer particles and thus to estimate the mean transport of the bedload (table 3), we used the empirical relation of Church and Hassan [1992]. The Church and Hassan [1992] relation was not developed for step-pool streams, but we chose this relation because it has also been tested on, and is generally supported by, field data from mountain streams [Scheingross et al., 2013], and on average it fits the Erlenbach data well (see section 3.1.2). The Church and Hassan [1992] relation expresses the scaled transport distances ($L^*$) of bedload particles as a function of the scaled grain sizes

$$L^* = L_i / L_{D50Surf} = 1.77 \left[ 1 - \log_{10} \left( D_i / D_{50Sub} \right) \right]^{1.35}$$

(2)

Where $L_i$ is the transport distance of individual grains of diameter $D_i$, $L_{D50Surf}$ is the mean transport distance of the median surface grain size ($D_{50Surf}$) and $D_{50Sub}$ is the median grain size of the stream bed subsurface. The use of equation 1 to estimate transport distances for unmeasured grain size fractions is described in the supplementary material, section S3.

2.3.2 Rio Cordon

In Rio Cordon, the transport distances of magnetically tagged tracer stones along the stream were measured during individual snowmelt and flood events from 1993 to 1998 (table 3) [Lenzi, 2004; Lenzi et al., 2006]. Two groups of 430 natural pebbles, cobbles, and boulders (32<$D<$512 mm) were painted and placed in May 1993 across two cross sections, in transverse rows with 1 to 2 m spacing. Movements of the marked clasts were mapped during repeated surveys (three transport events) from July 1993 to September 1994. Field measurements of displacement distances include the effects of a high-magnitude flood that occurred in September 1994. After this large flood, similar sets of two groups of 430 tracers were relocated on the previous two cross sections, and mapped from May 1996 to October 1998 (four transport events) [Lenzi, 2004]. Recovered percentages of marked clasts varied from 52 to 100%, depending on the flood magnitude.

The Rio Cordon tracer population matched the bed surface grain size distribution more closely than the Erlenbach tracer population did. Therefore it was unnecessary to estimate the transport distances of size fractions that were not represented by the tracer particles, as we did for the Erlenbach (see section 2.3.1.3.).
Bedload scaling relationships

Table 3: Bedload and transport distance characteristics for selected transport events *

<table>
<thead>
<tr>
<th>ID</th>
<th>Survey</th>
<th>Date</th>
<th>$Q_p$ [m³/s]</th>
<th>$\omega_{p-oc}$ [W/m²]</th>
<th>$\Sigma(\omega-oc)$ [J/m²]</th>
<th>$F_B$ [m³]</th>
<th>$f_B$ [m]</th>
<th>$L_T$ [m]</th>
<th>$L_B$ [m]</th>
<th>$S_{E_{LT}}$</th>
<th>$S_{E_{LB}}$</th>
<th>$h_B$ [m]</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1</td>
<td>18.06.09</td>
<td>3.0</td>
<td>939</td>
<td>5.46E+06</td>
<td>75</td>
<td>21.6</td>
<td>92</td>
<td>120</td>
<td>7.6</td>
<td>9.8</td>
<td>0.20</td>
</tr>
<tr>
<td>2</td>
<td>07.07.09</td>
<td>5.0</td>
<td>1534</td>
<td>6.37E+06</td>
<td>3.34E+06</td>
<td>123</td>
<td>37.9</td>
<td>143</td>
<td>186</td>
<td>14.5</td>
<td>18.9</td>
<td>0.23</td>
</tr>
<tr>
<td>4</td>
<td>21.07.09</td>
<td>1.3</td>
<td>363</td>
<td>6.37E+06</td>
<td>1.95E+06</td>
<td>43</td>
<td>12.3</td>
<td>56</td>
<td>73</td>
<td>18.1</td>
<td>23.5</td>
<td>0.20</td>
</tr>
<tr>
<td>5</td>
<td>26.07.09</td>
<td>1.3</td>
<td>370</td>
<td>2.74E+06</td>
<td>1.95E+06</td>
<td>26</td>
<td>7.5</td>
<td>37</td>
<td>48</td>
<td>9.8</td>
<td>12.7</td>
<td>0.18</td>
</tr>
<tr>
<td>6</td>
<td>04.08.09</td>
<td>1.2</td>
<td>321</td>
<td>3.34E+06</td>
<td>1.95E+06</td>
<td>10.7</td>
<td>3.6</td>
<td>27</td>
<td>35</td>
<td>6.3</td>
<td>8.2</td>
<td>0.10</td>
</tr>
<tr>
<td>7</td>
<td>12.08.09</td>
<td>1.3</td>
<td>375</td>
<td>1.95E+06</td>
<td>6.76E+05</td>
<td>0.1</td>
<td>0.03</td>
<td>6.9</td>
<td>9</td>
<td>0.9</td>
<td>1.2</td>
<td>0.004</td>
</tr>
<tr>
<td>8</td>
<td>22.06.10</td>
<td>0.5</td>
<td>85</td>
<td>1.07E+05</td>
<td>1.85E+05</td>
<td>0.4</td>
<td>0.12</td>
<td>14</td>
<td>18</td>
<td>2.3</td>
<td>3.0</td>
<td>0.01</td>
</tr>
<tr>
<td>9</td>
<td>12.07.10</td>
<td>0.7</td>
<td>165</td>
<td>1.85E+05</td>
<td>6.76E+05</td>
<td>0.4</td>
<td>0.11</td>
<td>14</td>
<td>18</td>
<td>8.3</td>
<td>10.8</td>
<td>0.01</td>
</tr>
<tr>
<td>11</td>
<td>19.07.10</td>
<td>0.8</td>
<td>168</td>
<td>6.76E+05</td>
<td>2.54E+06</td>
<td>11.7</td>
<td>3.3</td>
<td>29</td>
<td>38.0</td>
<td>9.7</td>
<td>12.6</td>
<td>0.11</td>
</tr>
<tr>
<td>12</td>
<td>26.07.10</td>
<td>1.6</td>
<td>467</td>
<td>2.54E+06</td>
<td>3.24E+06</td>
<td>31.3</td>
<td>9.0</td>
<td>36</td>
<td>47</td>
<td>3.6</td>
<td>4.7</td>
<td>0.22</td>
</tr>
<tr>
<td>17</td>
<td>25.08.94</td>
<td>2.0</td>
<td>635</td>
<td>3.24E+06</td>
<td>1.45E+07</td>
<td>375</td>
<td>107</td>
<td>161</td>
<td>209</td>
<td>27.4</td>
<td>35.6</td>
<td>0.57</td>
</tr>
<tr>
<td>20</td>
<td>28.07.95</td>
<td>10</td>
<td>2852</td>
<td>1.45E+07</td>
<td>1.45E+07</td>
<td>375</td>
<td>107</td>
<td>161</td>
<td>209</td>
<td>27.4</td>
<td>35.6</td>
<td>0.57</td>
</tr>
</tbody>
</table>

*Flood event number (ID); peak discharge ($Q_p$); unit excess stream power of peak discharge ($\omega_{p-oc}$); cumulative unit excess stream energy ($\Sigma(\omega-oc)$); total bedload volume ($F_B$); unit bedload volume ($f_B$); mean tracer transport distance ($L_T$); mean transport distance estimated for the total bedload ($L_B=130\%$ of $L_T$).
Standard error ($S_{E_{LT}}$) for the mean tracer transport distances; Standard error for the total transport distances of the total bedload ($S_{E_{LB}}=130\%$ of $S_{E_{LT}}$); Average active layer depth inferred from bedload volume and average transport distance ($h_B$). Rio Cordon events (RC1-RC7).

2.4. Bedload transport measurements

2.4.1 Erlenbach: The PBIS/Geophone system

Sediment transport has been monitored continuously in the Erlenbach since 1986 using piezoelectric bedload impact sensors (PBIS), and since 2000 using geophone sensors. The PBIS/Geophone system is a well-established indirect measurement method that has been used in several streams to study sediment transport. For technical details of the system see supplementary material, section S4 and Rickenmann and McArdell [2007], Rickenmann et al. [2012], Turowski et al. [2009], Turowski and Rickenmann [2011] and Turowski et al. [2011].

The geophone data are calibrated against sediment volumes accumulated in the retention basin (including fine material and pore volume). Therefore we estimated bedload transport by subtracting the percentage of fine material that was likely to be transported in suspension rather than as bedload, based on the approaches of Dade and Friend [1998] and Wu and Wang [2006] (see supplementary material, section S4). Based on the calibrated geophone data, we determined total bedload volumes for the transport events in which we measured tracer transport distances (table 3). From the total bedload volumes and stream width, we estimated the bedload volume per unit width, termed ‘unit bedload volume’.
2.4.2 Rio Cordon: Grid and storage basin

Sediment loads were determined by using ultrasonic sensors to survey volumetric changes in sediment deposits in the storage basin at the downstream end of the main study reach [e.g., Lenzi et al., 2004]. There, an inclined grid allowed the separation of coarse sediment ($D > 20$ mm) from water and fine sediment.

3. RESULTS

3.1. Tracer movements

3.1.1 Influence of flow magnitude on the transport distances (Erlenbach only)

Flow magnitude, as expressed by peak discharge, had a strong influence on the distribution of individual particles' transport distances (figure 4). The right-hand tail of the transport distance distribution dropped off quickly for low-magnitude events (figure 4 a-f), but was much longer for events with peak discharges larger than 3 m$^3$/s (figure 4 j-l). A gamma distribution was fitted to the tracer transport distance distribution for better visualization. The limited number of data points and the possible truncation of the distributions due to unrecovered far-moving particles (so-called "front-runners") prevent us from doing more detailed statistics on the transport distance distributions. However, the right-skewed and thin-tailed distributions that we observed, at least during low- to moderate-magnitude events, are also supported by literature results [Hassan et al., 2013; Liébault et al., 2012]. From these transport distance distributions, we infer that there are probably few, if any, unrecovered front-runners that would significantly change the calculated mean transport distances for low- and moderate-magnitude events. For the high-magnitude flood events with a peak discharge larger than 3 m$^3$/s, we assume that the derived mean transport distances are underestimated due to unrecovered front-runners that were flushed out of the study reach.

3.1.2 Fractional transport distances (Erlenbach only)

The transport distances of individual grains were also influenced by their particle size, especially during low- and moderate-magnitude events. Figure 5a shows the average transport distances of tracer particles grouped into five logarithmic grain size bins with centers ranging from 33.6 to 134.5 mm (spanning the tracer grain size range 28 mm and 160 mm, respectively), and grouped into three event size classes according to peak discharge ($Q_p < 1.5$ m$^3$/s, $Q_p 1.5 - 3$ m$^3$/s, and $Q_p > 3$ m$^3$/s, representing low, moderate, and high peak discharges). For the low- and moderate-magnitude events, the mean transport distance decreases with increasing particle size, whereas the transport distances during high flow events are almost unaffected by particle size (figure 5a).

In figure 5b, the transport distances of each individual particle ($L_{Si}$) are shown, scaled by the transport distance $LSD_{D50}$ of the size fraction that contains the median surface grain size $D_{50Surf}$; these scaled transport distances are compared to scaled grain sizes ($D_i$ scaled by the median grain size $D_{50Sub}$ of the streambed subsurface). There is obvious scatter in the individual scaled transport distances, but their medians (figure 5b) generally follow the pattern of the Church and Hassan [1992] equation (equation 1).
Figure 4: Tracer transport distance histograms for selected transport events sorted from (a) to (l) according to peak discharge ($Q_p$), with fitted gamma distributions (heavy black lines). $N_m$ is the number of moved tracers that the histograms are based on.

This justifies using equation 1 to estimate fractional transport distances from measured tracer transport distances. From the Church and Hassan [1992] equation, we estimated that the uncertainty in the mean transport distance due to unrepresentatively sampled grain sizes within the tracer population is smaller than 10% of the mean observed transport distance (see also supplementary material section S3). From the mean transport distances of the tracer population ($L_T$) and the tracer and bedload grain size distributions (figure 2) we estimated the mean transport distance ($L_B$) of the total bedload ($F_B$) to be 30% higher than the mean tracer transport distance (equation 1 and equations S2b & S2c, supplementary material section S3). The difference in transport distances arises primarily from the finer grain sizes (with generally longer transport distances) that make up a substantial proportion of the total bedload, but are absent from the tracer particle distribution.
Figure 5: (a) Mean transport distances, with standard errors, for three peak discharge classes \(Q_p<1.5\) m\(^3\)/s, \(1.5\leq Q_p<3\) m\(^3\)/s, and \(Q_p\geq3\) m\(^3\)/s, and five particle size classes indexed by their geometric means (33.6-134.5 mm), plotted as a continuous variable on a log scale. (b) Relationship between scaled travel distance \((L^*)\) and scaled particle size \((D^*)\) for all individual particle movements, and their medians (circles) in five grain-size classes. The transport distance \(L_s\) of each particle is scaled by the mean displacement of the grain size fraction containing the \(D_{50\text{Surf}}\) of the streambed surface; \(D_i\) is scaled by the \(D_{50\text{Sub}}\) of the subsurface. The thin black line (CH1992) represents equation 1 with the uncertainty bounds (dashed lines) given by Church and Hassan [1992].

3.2. Scaling mean transport distance with stream power/energy

The mean transport distance \((L_B)\) of the low- to moderate-magnitude events scaled roughly linearly with the excess stream power at peak discharge (figure 6, table 4), and, with a somewhat stronger correlation, \(L_B\) scaled roughly as the square root of the cumulative excess stream energy (figure 6b). In addition to the low- to moderate-magnitude events (those with a peak discharge < 3m\(^3\)/s), the high-magnitude flood events (those with a peak discharge > 3m\(^3\)/s) are also shown (light gray circles in figure 6). The transport distances of the low-, moderate- and high-magnitude events exhibit a consistent scaling relationship with peak stream power, with a slope of approximately 1 (figure 6a).

As a function of cumulative stream energy (figure 6b), transport distances for these high-magnitude events plot somewhat above the regression line of the low- to moderate-magnitude events, resulting in a significantly steeper scaling exponent for the fitted moderate- to high-magnitude events. (Note also that the transport distances during the high-magnitude events may be underestimated due to unrecovered front-runners). As a function of peak stream power, transport distances for the low-magnitude events at the Rio Cordon lie somewhat below the Erlenbach regression line, and the general trend appears to be slightly steeper (figure 6a, table 4). Compared to literature results, the transport distances observed at the Erlenbach and Rio Cordon are shorter than those observed at the Lainbach, Germany [Gintz et al., 1996], Carnation Creek, Canada [Haschenburger, 2013; Haschenburger and Church,
Bedload scaling relationships

1998] and several Ardennian rivers, Belgium [Houbrechts et al. 2012] (figure 6), which all have gentler channel gradients. For more details on the literature data and its treatment, see supplementary material section S5 and table S1.

Although the transport distances of the Erlenbach and the Rio Cordon tend to be generally lower compared to the literature results, the slopes of the Erlenbach and Rio Cordon regression lines in figure 6a and 6b are in a similar range as most of the power-law slopes fitted for other streams (table 4). Only for the Ardennian river dataset, which includes data from different streams, is the fitted power law characterized by a lower exponent (but also a lower r²).

3.3. Scaling unit bedload volume with stream power/energy

The transported unit bedload volumes of all observed transport events in the Erlenbach scale roughly as the cube of the peak stream power (table 4). The low- to moderate-magnitude events exhibit even steeper scaling, with a power law exponent of 3.85 (figure 7a, table 4). The unit bedload volumes scale roughly as the square of the cumulative stream energy (figure 7b) for the moderate to high-magnitude events and as a bit less than the square for the low- to moderate-magnitude events.

Figure 6: Mean transport distances (L_B), related to (a) the excess stream power of the peak discharge, and (b) the cumulative excess stream energy over each entire flood event. Black, dark gray and light gray circles indicate survey periods with low, moderate and high peak discharges (Q_p) at the Erlenbach (EB) (low: \(Q_p < 1.5 \text{ m}^3/\text{s}\); mod: \(1.5 \leq Q_p < 3 \text{ m}^3/\text{s}\); high: \(Q_p \geq 3 \text{ m}^3/\text{s}\)). Small symbols show initial displacements after particle seeding; large symbols show subsequent displacements from ‘natural’ positions for the Erlenbach. Error bars indicate standard errors. The black regression line is derived from functional analysis fitted to the low- to moderate-magnitude events, corrected for log-transformation bias. The dashed gray regression line is fitted to moderate- to high-magnitude events using the same methods. In (a), mean transport distances at Rio Cordon (RC), Carnation Creek [Haschenburger and Church, 1998], the Lainbach [Gintz et al., 1996] and for several Ardennian Rivers (Ard. Rivers) [Houbrechts et al., 2012] are also shown. In (b) the relation of Haschenburger[2013] for the Carnation Creek is added.
Figure 7: Unit bedload volumes ($f_B$) of the Erlenbach (EB), Rio Cordon (RC) and the Ardennian rivers (Ard. Rivers) related to (a) the excess stream power of the peak discharge; (b) the cumulative excess stream energy over each entire flood event. Black, dark gray and light gray circles indicate survey periods with low, moderate and high peak discharges ($Q_p$) at the Erlenbach. EB errors for $F_B$ are assumed to be a factor of two [Rickenmann and McArdell, 2008]. Regression lines were derived from the Erlenbach values by functional analysis.

Figure 8: Unit bedload volumes ($f_B$) as a function of mean transport distance ($L_B$) for the Erlenbach (EB), Rio Cordon (RC) and the Ardennian rivers (Ard. Rivers). Black, dark gray and light gray circles indicate events with low, moderate and high magnitude peak discharges ($Q_p$) at EB. EB errors for $F_B$ are assumed to be a factor of two [Rickenmann and McArdell, 2008] and error bars for $L_B$ show the standard error of the transport distances. The black regression line is fitted to the EB low- to moderate-magnitude events, and the dashed gray regression line is fitted to the moderate- to high-magnitude events.
The unit bedload volumes are more closely related to cumulative stream energy than to the peak stream power, at least for the low- to moderate-magnitude events. Unit bedload volumes of the Ardennian rivers are generally higher than those observed at the Erlenbach and Rio Cordon under comparable stream power conditions (figure 7a).

The unit bedload volume scales roughly as the square of the mean tracer transport distance in the Ardennian rivers, but roughly as the cube in the low- and moderate-magnitude Erlenbach events, and in Rio Cordon (figure 8 and table 4). For the Erlenbach, the higher-magnitude events (with peak discharges $Q_p \geq 3$ m$^3$/s) plot somewhat to the right of the regression line defined by events with $Q_p < 3$ m$^3$/s, although the mean transport distances are likely to be underestimated (and thus one would expect them to plot even further to the right).

4. DISCUSSION

Both the mean transport distances and the unit bedload volumes of our steep mountain streams, Erlenbach and Rio Cordon, are lower compared to data from other streams (Figure 6 and 7). However, their dependence on peak discharge, and cumulative stream energy are broadly consistent with the patterns observed for the other streams, except for the Ardennian data set. The Ardennian dataset may deviate because it consists of four different streams with only few data points per stream (one to four). This could explain the lower power-law slopes and weak correlations of mean transport distances and unit bedload volumes against stream power.

The generally shorter transport distances in the Erlenbach and Rio Cordon compared the other streams for a given excess peak stream power might be explained by the different stream types. Whereas Carnation Creek and Lainbach are lower-gradient mountain streams (stream gradient 0.01 m/m and 0.02 m/m, respectively) and the Ardennian data set also includes larger lowland rivers (gradients of 0.001-0.011), the mountain streams Erlenbach and Rio Cordon are characterized by steep channels (gradients of 0.17 and 0.13, respectively), step-pools and rough bed topography. The rough bed topography might be responsible for relatively lower bedload volumes and transport distances, compared to the other sites. Bedload transport is relatively lower in steep streams compared to lowland rivers, due to a reduced transport efficiency resulting from an increased demand for particle entrainment [e.g., Bathurst, 2013; Bunte et al., 2013; Lamb et al., 2008; Mueller et al., 2005] and a reduced energy available for sediment transport resulting from macro-roughness [e.g., Bathurst, 1987; Nitsche et al., 2011; Chiari & Rickenmann, 2011; Yager et al., 2012a].

The cumulative stream energy, which represents the integrated energy of an entire flow event, has a larger dynamic range (i.e., varies over more orders of magnitude) than peak stream power does. Perhaps partly for this reason, it also appears to be somewhat better than peak stream power at explaining the variation in transport distance and unit bedload volume among low- and moderate-magnitude events at the Erlenbach (figure 6, figure 7, and table 4). Haschenburger [2013] also found a very strong correlation ($r^2 = 0.99$) between mean transport distance and cumulative stream energy for Carnation Creek.
### Chapter IV

Table 4: Fitted power laws (y = αx^β)

<table>
<thead>
<tr>
<th>Dataset</th>
<th>x=Peak stream power</th>
<th>x=Cumulative stream energy</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>α</td>
<td>β</td>
</tr>
<tr>
<td>EB low-mod</td>
<td>0.06</td>
<td>1.12</td>
</tr>
<tr>
<td>EB mod-high</td>
<td>0.05</td>
<td>1.1</td>
</tr>
<tr>
<td>EB all</td>
<td>0.14</td>
<td>0.97</td>
</tr>
<tr>
<td>RC</td>
<td>0.02</td>
<td>1.25</td>
</tr>
<tr>
<td>Carn. Creek</td>
<td>0.005</td>
<td>1.72</td>
</tr>
<tr>
<td>Lainbach</td>
<td>0.14</td>
<td>1.37</td>
</tr>
<tr>
<td>Ard. Rivers</td>
<td>8.61</td>
<td>0.45</td>
</tr>
</tbody>
</table>

**y=Transport distance**

<table>
<thead>
<tr>
<th>Dataset</th>
<th>x=Mean transport distance</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>α</td>
</tr>
<tr>
<td>EB low-mod</td>
<td>5.7E-10</td>
</tr>
<tr>
<td>EB mod-high</td>
<td>4.6E-05</td>
</tr>
<tr>
<td>EB all</td>
<td>2.3E-07</td>
</tr>
<tr>
<td>RC</td>
<td>3.3E-12</td>
</tr>
<tr>
<td>Ard. Rivers</td>
<td>5.7E-03</td>
</tr>
</tbody>
</table>

**y=Unit bedload volume**

<table>
<thead>
<tr>
<th>Dataset</th>
<th>x=Mean transport distance</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>α</td>
</tr>
<tr>
<td>EB low-mod</td>
<td>1.3E-08</td>
</tr>
<tr>
<td>EB mod-high</td>
<td>7.6E-04</td>
</tr>
<tr>
<td>EB all</td>
<td>1.8E-06</td>
</tr>
<tr>
<td>RC</td>
<td>2.6E-11</td>
</tr>
<tr>
<td>Carn. Creek</td>
<td>6.7E-05</td>
</tr>
<tr>
<td>Ard. Rivers</td>
<td>7.2E-03</td>
</tr>
</tbody>
</table>

**y=Active layer**

<table>
<thead>
<tr>
<th>Dataset</th>
<th>x=Mean transport distance</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>α</td>
</tr>
<tr>
<td>EB low-mod</td>
<td>8.7E-06</td>
</tr>
<tr>
<td>EB mod-high</td>
<td>7.2E-03</td>
</tr>
<tr>
<td>EB all</td>
<td>6.1E-05</td>
</tr>
<tr>
<td>RC</td>
<td>1.1E-04</td>
</tr>
<tr>
<td>Ard. Rivers</td>
<td>8.1E-04</td>
</tr>
</tbody>
</table>

---

However, transport distances of the high-magnitude Erlenbach events clearly deviate from the cumulative stream energy relationship defined by the low- and moderate-magnitude events (figure 6); if the assumed underestimation of these transport distances (due to potentially unrecovered front-runners) could be corrected, this deviation would presumably be even larger.

### 4.1. Back-calculation of an active layer depth

In our Erlenbach and Rio Cordon data, bedload volumes scale much more steeply than transport distances do, as functions of peak stream power (power-law slopes of roughly 3 and 1, respectively). Bedload volumes also scale much more steeply with cumulative stream energy than transport distances do (power-law slopes of roughly 1.8 and 0.5, respectively).
These observations imply that large bedload volumes during large storms may arise primarily from deeper and wider bed scouring, and only secondarily from longer transport distances. The impact of the water flow on the streambed can be discussed using the active layer concept. In this concept, as a modeling assumption, the probability distribution for a particle to be entrained by the flow is simplified to a step-function dividing the sediment vertically into two layers [Hirano, 1971; Parker, 1991; Parker et al., 2000]: an immobile layer underlying a temporally and spatially variable near-surface active layer, in which the probability of erosion per unit time is the same for every grain. In this study we consider the active layer as a virtual layer in which all of the particles are entrained and which is mobilized over the entire stream width during the course of the transport event.

From measurements of transport distances ($L_B$), bedload volumes ($F_B$), and average stream width ($w$), we estimated an active layer depth ($h_B$), describing the flood impact on the streambed, for individual transport events (figure 9) via the sediment continuity equation:

$$h_B = \frac{F_B}{L_B w}$$

The mean derived active layer thickness ($h_B$) that we calculated from equation 2 for the Erlenbach is about 0.01-0.22 m for the low- to moderate-magnitude events and about 0.57 m for the largest transport event observed (table 3), which is roughly similar to the $D_{90\text{Surf}}$ of the bed surface. Since the mean transport distances are potentially underestimated for the high-magnitude events, the back-calculated active layer depths (light grey circles, figure 9) might be overestimated for these events. Although the high values might be somewhat overestimated, these values of the active layer depth are similar to independent measurements from the Erlenbach. Consider, for example, the active layer depth of 0.57 m inferred for event 20, which had a peak discharge of 10 m$^3$/s. For an event with a similar peak discharge (9.3 m$^3$/s; event 15, table 2) pre- and post-flood long-profile measurements show that the mean erosion was 0.47 m along a 600-m channel reach, and the maximum local erosion was 2.8 m [Turowski et al., 2013]. In addition, maximum mobile grain sizes of around 0.5 m were observed in another comparable transport event [Turowski et al., 2009].

Inferred active layer depths in the Erlenbach scale approximately linearly with cumulative stream energy, for both large and small events (figure 9b). The relationship between active layer depth and peak stream power, by contrast, cannot be described by a single power function over the entire range of flow magnitudes (figure 9a). The data shown in figure 9a indicate two trends: one relation with a power-law slope of 2.7 for low-to-moderate magnitude events, and a second relation with a power-law slope of 0.8 for events with moderate-to-high flow magnitudes. The trends in active layer depth in figure 9a are similar to those shown in figure 7a for unit bedload volumes. Both are steep functions of peak stream power for low-to-moderate flow magnitudes, and markedly shallower functions of peak stream power for moderate-to-high flow magnitudes. This shift could arise if, particularly at low or moderate flows, only a fraction of the channel width actively transports sediment [Dietrich et al., 1989], and if this active layer width increases with increasing flow.
Chapter IV

Figure 9: Back-calculated active layer depth related to unit excess stream power of the peak discharge (a) and unit cumulative excess stream energy (b) for the Erlenbach (EB) and Rio Cordon (RC). Black, dark gray and light gray circles indicate events with low, moderate and high peak discharges ($Q_p$) at the Erlenbach (EB). The solid black lines were fitted to the Erlenbach low- to moderate-magnitude events ($Q_p < 3 \text{ m}^3/\text{s}$). The dashed black line in (a) was fitted to the moderate- to high-magnitude events ($Q_p \geq 1.5 \text{ m}^3/\text{s}$). Erlenbach error bars are derived from uncertainties in the measured bedload volume (factor 2) and the standard error of the transport distances. Squares represent scour chain measurements in Carnation Creek, Canada [Haschenburger and Church, 1998]. Small black dots (Ard. Rivers) are based on scour chain measurements in several Ardennian rivers, Belgium [Houbrechts et al., 2012]. The black and dashed gray lines were not fitted to data from Rio Cordon, Carnation Creek, or the Ardennian rivers.

Thus at low-to-moderate flows, the active layer width, active layer depth, and average transport distance may all increase with increasing flow, with the result that bedload transport is a steep function of flow magnitude. One may further hypothesize that in moderate and larger events, the active layer width reaches both banks, and thus cannot increase further; thus only active layer depth and transport distance can continue to increase, making bedload transport a shallower function of stream power in this magnitude range.

To understand how this hypothesized mechanism might affect the active layer depths shown in figure 9a, one must remember that these are virtual quantities averaged over the entire channel width. They represent the actual active layer depth, times the ratio between the actual active layer width and the channel width. Thus, the inferred active layer depth would reflect the increase in the real-world width and depth of the active layer in the low-to-moderate magnitude range (and thus be a relatively steep function of stream power), but only the increase in real-world depth of the active layer at higher flow magnitudes (where the real-world active layer width has maxed out at the full width of the channel).

Although the total active layer depth in the Erlenbach can be related to both cumulative stream energy and peak stream power (figure 9), peak stream power provides a basis for comparison with other field observations, namely at Rio Cordon, Carnation Creek...
Bedload scaling relationships

[Haschenburger and Church, 1998] and the Ardennian rivers [Houbrechts et al., 2012], where cumulative stream energy could not be estimated. The estimated active layer depths at Rio Cordon (white diamonds, figure 9a) follow a similar trend to the Erlenbach observations during moderate-high flow magnitudes (solid black circles, figure 9a), although with more scatter. The two very low active layer depth values obtained in Rio Cordon, lying below the Erlenbach trend lines (figure 9a), were observed under supply-limited conditions before the September 1994 flood event as reported by Lenzi [2004]. The 1994 flood was an exceptional event which broke up the stable bed structures and led to almost unlimited sediment supply [Lenzi et al., 2004]. The bedload volumes (and thus active layer depths) observed at Rio Cordon are markedly larger after this extreme flood event.

The Ardennian and Carnation Creek active layer depths shown in figure 9a were measured by scour chains. They follow a similar trend as the Erlenbach back-calculated active layer depths for moderate-high flow magnitudes, but lie well above the Erlenbach values for low-magnitude stream flow magnitudes. The smaller active layer depths in the Erlenbach compared to the Ardennian rivers and Carnation Creek during low-magnitude flow conditions (figure 9a) might be explained by the bed configuration of steep mountain streams with more stable bed structure (form and grain roughness, bedrock, woody debris), and a more limited sediment supply, compared to lower gradient streams. Alternatively, the smaller inferred active layer depth in the Erlenbach may be an artifact of changes in the width of the active layer, as outlined in the hypothesis above. A third possible explanation might be the different measurement techniques used. Our back-calculation methods average the active layer depth over space and time, whereas scour chains measure a spatially local, and temporally maximal, active layer depth for a given transport event.

5. CONCLUSIONS

We have presented measurements of bedload transport volumes and transport distances of tracer particles of two steep mountain streams, the Erlenbach (Switzerland) and Rio Cordon (Italy). Despite low tracer recovery rates at the Erlenbach (particularly for large events) and the resulting limited number of data points (table 2), the observed transport distances and unit bedload volumes were strongly correlated to both the peak stream power and the cumulative stream energy (figures 6 and 7) of individual transport events.

Both the Erlenbach and Rio Cordon are characterized by shorter transport distances and smaller unit bedload volumes than those observed in several lower-gradient streams (figures 6 and 7). However, transport distance and unit bedload volumes in all of these streams exhibit broadly similar scaling with peak stream power (table 4), implying that these scaling relationships are not strongly dependent on stream gradient.

Unit bedload volumes scale much more steeply than tracer transport distances do, as functions of either stream power or cumulative stream energy (figures 6 and 7). Furthermore, unit bedload volumes scale as roughly the third power of transport distances (figure 8). These observations imply that storm-to-storm variations in bedload volumes arise predominantly through variations in scour depth and width rather than through variations in transport distances. From the observed bedload volumes and transport distances, we back-calculated an effective mean active layer depth, which also was strongly correlated with both peak stream
power and cumulative stream energy (figure 9), and was broadly consistent with independent measurements of scour depth.

Approximating bedload transport in steep, rough mountain streams by a continuous, uniform active layer involves gross simplifications, which will be challenged by (i) partial sediment transport [Wilcock and McArdell, 1997], (ii) mobile and stationary zones of the streambed [e.g., Hassan et al., 2005; Marquis and Roy, 2012; Yager et al., 2012b], (iii) variations in the horizontal and vertical dimensions of the active layer over time [e.g., Haschenburger, 1996; Hassan, 1990; Wilcock et al., 1996], and (iv) variations in the fraction of the channel width that is actively transporting sediment [Dietrich et al., 1989]. Despite these obvious limitations, the active layer concept provides a useful simplified characterization of streambed response to hydraulic forcing during bedload transport. Independent, direct measurements of active layer depths in steep streams would be helpful for interpreting the active layer depths results derived from the back-calculation procedure presented here.

ACKNOWLEDGEMENTS

This study was supported by CCES project APUNCH and SNF grant 200021_124634/1, the Swiss Federal Research Institute WSL and the Swiss Federal Institute of Technology ETH. We thank K. Bunte for providing the Erlenbach grain-size distribution data. We thank A. Badoux, B. J. MacVicar, the editor and several anonymous reviewers for their constructive remarks on the manuscript.

REFERENCES


Bunte, K., and S. R. Abt (2001b), Sampling surface and subsurface particle-size distributions in wadable gravel-and cobble-bed streams for analyses in sediment transport, hydraulics,
Bedload scaling relationships

and streambed monitoring, 428 pp, U.S. Department of Agriculture, Forest Service, Fort Collins, CO.


Chapter IV

Wiley & Sons Ltd.
Klingeman, P. C. (1979), Sediment transport research facilities, Oak Creek, OregonRep., Water Resources Institute, Oregon State University, Corvallis.
Laronne, J. B., D. N. Outhet, J. L. Duckham, and T. J. McCabe (1992), Determining event bedload volumes for evaluation of potential degradation sites due to gravel extraction, N.S.W., Australia, 87-94 pp, International Association of Hydrological Sciences,, Wallingford, Oxfordshire.
Lenzi, M. A. (2004), Displacement and transport of marked pebbles, cobbles and boulders during floods in a steep mountain stream, *Hydrological Processes*, 18(10), 1899-1914, doi:


Chapter IV


Schwer, P., and D. Rickenmann (1999), Single particle movements in steep torrents: Observations with magnetically tagged tracer stones, WSL, Birmensdorf, Switzerland.


Bedload scaling relationships


CHAPTER V

CONCLUSIONS

The outcomes of the studies conducted within this dissertation added to the physical understanding of transport processes, i.e. the influence of macro-roughness at steep gradients on flow velocity and bedload transport in steep mountain streams. The results might be used for prediction of flow velocity and bedload transport, which is important in various fields such as aquatic ecology, evolution of the landscape, water management and natural hazard prediction.

Study I addressed the question how the channel bed, i.e. the bed grain size distributions or bed roughness heights, and flow velocity adjust to each other at increasing bed gradients. The mountain stream presented in this study can be considered as a natural experiment, where bed gradient increases over roughly one order of magnitude along the 1-km long study reach, while flow discharge and width remain approximately constant. We found that the characteristic grain size \( D_{84} \) is in close agreement to the statistically better supported but much more time- and cost consumptive TLS (terrestrial laser scanning) point cloud derived roughness measures. Also with respect to flow velocity prediction, using the \( D_{84} \) as a roughness height provided comparable results as when using the point cloud derived roughness measures. Relating measured flow velocity to measured bed roughness was performed using the Darcy-Weisbach relation (scaling flow depth) and the closely related concept of dimensionless hydraulic geometry (scaling unit discharge). Both concepts work generally well to explain and predict measured flow velocity at increasing bed gradients. However, better results were obtained for the concept of the hydraulic geometry. These observations are in close agreement with the findings previously published mainly for lower to medium gradient streams [e.g., Ferguson, 2007; Nitsche et al., 2012; Rickenmann and Recking, 2011; Yochum et al., 2012], and they generally justify the use of both concepts for flow velocity prediction at very steep slopes, for which to date little data have been reported.

The adjustment of the streambed, i.e. the streamed grain size distributions, to given flow conditions and bed gradient is closely related to the particle sizes that can be entrained by the flow. The critical shear stress can be seen as a proxy controlling the feedback system between bed gradient and morphology, flow conditions and bedload transport. Empirical critical shear stress approaches, incorporated in a flow resistance equation, provided good estimates of the \( D_{84} \) in comparison to the measured \( D_{84} \). Given that the empirical relations are derived from entirely independent data, on the one hand allows the conclusion that the method presented here is able to adequately predict \( D_{84} \) as a function of bed slope and flow conditions, and on the other hand provides further physical support for the presented empirical relations.

The first question of Study II asked, if accounting for the wide grain size distribution, typical for steep slopes, using fractional bedload transport models can improve bedload transport prediction as compared to models that compute bedload transport for a median grain size. It was found that bedload transport predictions do not significantly differ whether calculated per grain size fraction or for a median grain size. It appeared that potential hiding and exposure effects are relatively insignificant, although influencing the threshold of motion
Conclusions

of individual size classes, compared to the general uncertainties associated with a considerably reduced transport capacity at steep slopes. Furthermore, despite the broad range of stream characteristics investigated in this study, we could not identify a strong predictive parameter for the hiding exponent, which controls the degree of size selective transport.

Another focus of Study II was to show how two different approaches for bedload transport prediction can account for the reduced transport capacity at steep slopes, i.e. one only based on the increased flow-energy dissipation or one only based on the increased threshold for initiation of motion. It was shown that transport equations such as the Wilcock and Crowe [2003] equation, which were developed on flume data for lower stream gradients, can also be applied to steep mountain streams if one accounts for flow resistance due to macro-roughness and/or the increased demand of energy for initiation of motion. However, the choice of the one or the other approach has further implications for bedload transport predictions, as the stress-dependent increase in bedload transport rates strongly depends on which approach is used. Using bedload transport equations for bedload transport predictions therefore presumes knowledge under which assumptions the equation was derived, i.e. whether it is based on a macro-roughness derived reduced stress or an increased critical condition for initiation of motion.

Study III investigated how the movements of individual particles are related to bulk bedload volumes in two steep mountain streams, including a comparison with literature data. Both study streams, the Erlenbach and Rio Cordon, are characterized by shorter transport distances and smaller unit bedload volumes than those observed in several lower-gradient streams with generally finer and more homogeneous grain size distributions. This observation could indicate lower sediment availability and increased critical thresholds for initiation of motion. However, transport distance and unit bedload volumes in all of these streams exhibit broadly similar scaling with peak stream power implying that these scaling relationships are not strongly dependent on stream gradient. Bedload volumes scale much more steeply than tracer transport distances do, as functions of either stream power or cumulative stream energy. Furthermore, unit bedload volumes scale as roughly the third power of transport distances. These observations imply that storm-to-storm variations in bedload volumes arise predominantly through variations in scour depth and width rather than through variations in transport distances. From the observed bedload volumes and transport distances, we back-calculated an effective mean active layer depth, which also was strongly correlated with both peak stream power and cumulative stream energy, and which was broadly consistent with independent measurements of scour depth. This suggests that the active layer concept provides a useful simplified characterization of streambed response to hydraulic forcing during bedload transport.

The results of this dissertation clearly showed the influence of streambed roughness on both flow velocity and bedload transport rates. In Study I it could be observed that flow velocity does not increase with increasing bed gradients (for approximately constant discharge rates and flow width), even for bed gradients ranging over an order of magnitude from 3-40%. This could indicate that the kinetic energy acting on the grains should be roughly the same at all bed gradients. Furthermore, in Study I it could be shown that the flow resistance/flow velocity patterns observed in the Riedbach agree with the patterns observed in literature and
that the derived empirical flow resistance approaches are generally well suited for prediction of flow velocity and thus the kinetic energy. However, by considering bedload transport rates over wide ranges of bed gradients and related variations in stream bed roughness it could be shown [Study II] that for bedload transport prediction itself, a bed roughness related flow resistance approach, either based on the Manning/Darcy-Weisbach equations or the concept of dimensionless hydraulic geometry as validated in Study I is not sufficient to adequately predict bedload transport. Instead, systematic overestimations of bedload transport at steep bed gradients were observed [Study II] if the prediction is only based on a flow resistance equation. If the flow resistance equations validated in Study I is correct, then why are bedload transport rates lower at steeper bed gradients with increased macro-roughness under otherwise comparable hydraulic conditions?

To answer that question and demonstrate possible solutions for bedload transport predictions, a simple bedload transport equation is presented below,

\[ Q_s = k(Q - Q_c) \]  

(1)

where \( Q_s \) is the bedload transport rate, \( Q \) is the available flow energy (e.g., flow discharge, stream power or shear stress), \( Q_c \) is the critical condition for initiation of motion and \( k \) is a prefactor (more details see below). A common way to avoid overestimations of bedload transport rates is to further reduce \( Q \) based on roughness measures as e.g. a characteristic grain sizes [Rickenmann and Recking, 2011; Wilcock et al., 2009; Study II], but considering \( Q \) to be adequately represented by the flow resistance equations [Study I] this seems to come short of a full physical explanation. A second possibility is to enlarge the critical conditions \( Q_c \) at steeper stream reaches with higher bed roughness [e.g., Camenen, 2012; Lamb et al. 2008], based on empirical equations that explain a positively related \( Q_c \) with bed slope by ‘variable friction angles, grain emergence, flow aeration, changes to local flow velocity and turbulent fluctuations’ at steep slopes [e.g., Lamb et al., 2008; Prancevic et al., 2014].

However, there could be a third possible explanation for reduced bedload transport rates at steep slopes compared to low-land rivers under otherwise comparable hydraulic conditions. Besides generating resistance to the water flow, macro-roughness in steep streams could also generate resistance to the transported sediments and thus reduce the transport efficiency, captured by the prefactor \( k \) in equation (1). The assumption of a reduced transport efficiency at steep slopes is also supported by results from Study III, where lower transport distances of tracer particles at comparable hydraulic conditions (stream power) in two steep streams compared to lower gradient rivers were observed. Thus, in terms of the bedload transport equation presented above, transport predictions could be adjusted by reducing the transport efficiency described by the pre-factor \( k \) (Equation 1) to reduce the bedload transport overestimations that are typical for steep mountain streams.

It would be an interesting field for further research to quantify the effects of the reduced transport efficiency \( k \) on bedload transport rates, either based on experiments on bedload transport distances in mountain streams or laboratory flumes. In addition, further quantification of \( Q_c \) is necessary for improving the understanding of bedload transport processes. Despite a lot of previous work on the relations between \( Q_c \) and bed gradient [e.g., Diplas et al., 2008; Lamb et al., 2008; Prancevic et al., 2014; Prancevic and Lamb, subm], it is not fully clarified yet which part of the positive relation of \( Q_c \) and \( S \) is based on the increase...
in bed stability at steep slopes and which part is due to the increased resistance affecting bedload transport directly, and indirectly through the hydraulics.

REFERENCES


Prancevic, J. P., and M. P. Lamb (subm), Unraveling bed slope from relative roughness in initial sediment motion, Paper submitted to *Journal of Geophysical Research - Earth Surface.*


Wilcock, P. R., J. Pitlick, and Y. Cui (2009), *Sediment transport primer: estimating bed-material transport in gravel-bed rivers,* US Department of Agriculture, Forest Service, Rocky Mountain Research Station.
SUPPLEMENTARY MATERIAL: STUDY I (CHAPTER II)

Stream bed roughness and flow velocity in a steep mountain channel

AUTHORS

Johannes M. Schneider\textsuperscript{1,2}, johannes.schneider@wsl.ch (corresponding author)
Dieter Rickenmann\textsuperscript{1}, dieter.rickenmann@wsl.ch
Jens M. Turowski\textsuperscript{1,3}, turowski@gfz-potsdam.de
James W. Kirchner\textsuperscript{1,2}, kirchner@ethz.ch

1) Swiss Federal Research Institute WSL, Mountain Hydrology and Mass Movements, Zürcherstrasse 111, 8903 Birmensdorf, Switzerland
2) Swiss Federal Institute of Technology ETH Zürich, Institute of Terrestrial Ecosystems, 8092 Zürich, Switzerland
3) Helmholtz Centre Potsdam, GFZ German Research Centre for Geosciences, Telegrafenberg 14473 Potsdam, Germany

The auxiliary material includes supporting information on the Riedbach catchment characteristics, methodologies used as well as additional data.

TEXT S1.

Grain size distributions (GSD) were determined using the line-by-number (LBN) technique [Fehr, 1987], consisting of 290-1350 (median 380) counts of individual particles within the sampled reaches. To determine the surface GSD, the pebble count was transformed into a full GSD by assuming an average proportion of 10% fine material with $D<10$ mm, according to observations reported in Recking [2013] and in Anastasi [1984]. This assumption may be further justified for the Riedbach, for which a volume sample of the subsurface material in reach R#01 indicates a proportion of 13% to 28% of fine material with $D<10$ mm [Bunte et al., subm.]. Assuming a reduction of fine material by a factor of about 2, the surface layer would include a proportion of 6.5% to 14% of fine material with $D<10$ mm.

TEXT S2.

The assumption of a constant flow discharge along the 1 km study reach is supported by a comparison of the measured discharge rates at the water intake ($Q_{WI}$) with the discharge estimates derived from the dye tracer experiments ($Q_{FL}$) by the integration method (or gulp injection). We found that $Q_{FL}$ was generally larger than $Q_{WI}$, and the average ratio $Q_{FL}/Q_{WI}$ increased slightly, linearly in the downstream direction ($x$), from 1.2 at Fl#01 to 1.6 at FL#10.
Appendix

\(\frac{Q_{FL}}{Q_{WI}}=0.0003x+1.2\) with \(x\) representing distance in meters; \(R^2=0.63\) or roughly 25\% on average. However the general offset (factor 1.2) corresponds well to observed unmeasured residual water and overtopping water at high flows at the water intake. The increasing ratio \(\frac{Q_{FL}}{Q_{WI}}\) in the downstream direction is therefore mainly explained with increasing measurement uncertainties, partly due to light degradation of the fluorescein tracer (up to 5\%) and partly due to spreading of the BTC in the downstream direction, resulting in an increased amount of dye tracer that cannot be detected by the fluorometers. Both artifacts imply a stronger dilution of the dye tracer and thus overestimates of flow discharge.

REFERENCES


Figure S1: Distribution of daily peak discharges at the Riedbach water intake during the measurement period 2009 to 2014. Measurements were corrected for unmeasured residual and overtopping water by a factor of 1.2, valid for measured discharge rates up to about 4 m\(^3\)/s, corresponding to the maximum discharge that can be captured by (and measured in) the water intake. Extreme discharges with much more overtopping water due to intensive rainfall events could not be quantified and are not included in this figure.
Figure S2: Sediment availability in the recent glacier retreat area during low flows (Photo: Kristin Bunte)

Figure S3: Riedbach grain size distributions. PC refers to grid-by-number sampling (pebble count, blue dashed line). LBN refers to line-by-number sampling.
Appendix

Figure S4: Isotropic semi-variograms (black and gray dots) for selected illustrative sub-reaches with fitted spherical models (blue and gray lines) including definitions of the four semi-variogram characteristics range \( r \), nugget \( c_0 \), partial sill \( c_p \), and sill \( c \). Gray dots and lines show semi-variograms with less clearly defined ranges and sill levels. The blue and lower gray semi-variograms refer to sub-reaches within Reach R#01 (\( S=3\% \)), the upper gray semi-variogram refers to a sub-reach within R#03 (\( S=6\% \)).

Figure S5: a) Roughness height from meshed terrestrial laser scan point cloud data related to bed slope. b), c) and d) Roughness measures derived from meshed DEM in comparison to roughness measures derived from the point cloud directly.
Figure S6: Relationship between \((8/f_{at})^{0.5}\) and relative flow depth scaled by a) the \(IPR_{90}\) derived from line-by-number sampling and b) the standard deviation \(STD_z\) of the de-trended laser scan point clouds. Black thick lines show a power law fit to the data of all reaches; relationships for individual reaches (thin colored lines) are steeper. c) Dimensionless velocity related to dimensionless discharge scaled by \(IPR_{90}\) and d) by the \(STD_z\). In d) the equation given in Yochum et al. [2012] is shown.
Figure S7: Optimized parameter pair (magenta squares) of $a_1$ (left) and $a_2$ (right) defining the shape of the *Ferguson* [2007] equation, fitted to the Riedbach data based on 10,000 Monte Carlo simulations (blue dots). $a_1$ defines the part of the *Ferguson* (2007) equation for large relative flow depth ($d/R > 1-10$). Because the Riedbach data do not cover this range of relative flow depths, $a_1$ is not well constrained.

Figure S8: Relationship between Froude number and unit discharge for the 10 study reaches ($v=$velocity, $g=$gravitational acceleration, $d=$hydraulic radius, $S=$bed gradient).
### Table S1: Flow velocity measurements

<table>
<thead>
<tr>
<th>Date/Time</th>
<th>Inj.</th>
<th>R#01</th>
<th>R#02</th>
<th>R#03</th>
<th>R#04</th>
<th>R#05</th>
<th>R#06</th>
<th>R#07</th>
<th>R#08</th>
<th>R#09</th>
<th>R#10</th>
</tr>
</thead>
<tbody>
<tr>
<td>dd.mm.yy hh:mm</td>
<td>Q (m/s)</td>
<td>Loc</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>13.06.13 15:01</td>
<td>0.59</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.65</td>
<td>0.67</td>
<td>0.70</td>
</tr>
<tr>
<td>13.06.13 15:40</td>
<td>0.624</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.69</td>
<td>0.66</td>
<td>0.64</td>
</tr>
<tr>
<td>13.06.13 16:37</td>
<td>0.68</td>
<td>1</td>
<td>0.78</td>
<td>0.70</td>
<td>0.57</td>
<td>0.59</td>
<td>0.60</td>
<td></td>
<td>0.75</td>
<td>0.69</td>
<td>0.63</td>
</tr>
<tr>
<td>13.06.13 17:15</td>
<td>0.726</td>
<td>1</td>
<td></td>
<td></td>
<td>0.58</td>
<td>0.61</td>
<td>0.63</td>
<td></td>
<td>0.73</td>
<td>0.73</td>
<td>0.74</td>
</tr>
<tr>
<td>13.06.13 18:00</td>
<td>0.811</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.75</td>
<td>0.73</td>
<td>0.71</td>
</tr>
<tr>
<td>13.06.13 18:29</td>
<td>0.846</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.83</td>
<td>0.76</td>
<td>0.70</td>
</tr>
<tr>
<td>13.06.13 18:54</td>
<td>0.88</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.83</td>
<td>0.80</td>
<td>0.78</td>
</tr>
<tr>
<td>14.06.13 13:40</td>
<td>1.03</td>
<td>1</td>
<td>0.86</td>
<td>0.69</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.76</td>
<td>0.83</td>
<td>0.92</td>
</tr>
<tr>
<td>14.06.13 14:43</td>
<td>1.104</td>
<td>1</td>
<td>0.93</td>
<td>0.62</td>
<td>0.70</td>
<td>0.74</td>
<td></td>
<td></td>
<td>0.76</td>
<td>0.83</td>
<td>0.92</td>
</tr>
<tr>
<td>08.07.13 17:54</td>
<td>1.671</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.87</td>
<td>1.06</td>
<td>1.21</td>
</tr>
<tr>
<td>08.07.13 18:46</td>
<td>1.678</td>
<td>1</td>
<td>1.22</td>
<td>1.15</td>
<td>0.86</td>
<td>0.92</td>
<td>0.95</td>
<td>0.86</td>
<td>1.08</td>
<td>1.15</td>
<td>1.12</td>
</tr>
<tr>
<td>09.07.13 09:50</td>
<td>1.133</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.71</td>
<td>0.87</td>
<td>0.93</td>
</tr>
<tr>
<td>09.07.13 10:47</td>
<td>1.171</td>
<td>1</td>
<td>1.21</td>
<td>0.94</td>
<td>0.76</td>
<td>0.79</td>
<td>0.81</td>
<td>0.70</td>
<td>0.90</td>
<td>0.91</td>
<td>0.93</td>
</tr>
<tr>
<td>09.07.13 13:18</td>
<td>1.32</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.78</td>
<td>0.91</td>
<td>1.04</td>
</tr>
<tr>
<td>09.07.13 14:42</td>
<td>1.345</td>
<td>1</td>
<td>1.39</td>
<td>0.99</td>
<td></td>
<td>0.82</td>
<td></td>
<td>0.76</td>
<td>0.92</td>
<td>1.03</td>
<td>0.99</td>
</tr>
<tr>
<td>09.07.13 16:17</td>
<td>1.389</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.80</td>
<td>0.91</td>
<td>1.02</td>
</tr>
<tr>
<td>09.07.13 17:17</td>
<td>1.394</td>
<td>1</td>
<td>1.49</td>
<td>1.00</td>
<td>0.82</td>
<td>0.85</td>
<td>0.87</td>
<td>0.77</td>
<td>0.97</td>
<td>1.03</td>
<td>0.96</td>
</tr>
<tr>
<td>10.07.13 10:07</td>
<td>0.992</td>
<td>1</td>
<td>0.99</td>
<td>0.83</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>10.07.13 10:18</td>
<td>1.002</td>
<td>1</td>
<td>1.05</td>
<td>0.85</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>10.07.13 11:23</td>
<td>1.021</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.67</td>
<td></td>
<td></td>
</tr>
<tr>
<td>10.07.13 11:37</td>
<td>1.048</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.66</td>
<td></td>
<td></td>
</tr>
<tr>
<td>05.08.13 15:44</td>
<td>2.797</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.36</td>
<td></td>
<td></td>
</tr>
<tr>
<td>05.08.13 16:50</td>
<td>3.011</td>
<td>1</td>
<td>1.62</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.31</td>
<td></td>
<td>1.47</td>
</tr>
<tr>
<td>05.08.13 18:07</td>
<td>3.141</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>05.08.13 19:12</td>
<td>3.141</td>
<td>1</td>
<td>1.71</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.38</td>
<td></td>
<td>1.50</td>
</tr>
<tr>
<td>06.08.13 11:08</td>
<td>1.933</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.13</td>
<td>0.98</td>
<td>1.22</td>
</tr>
<tr>
<td>06.08.13 12:03</td>
<td>2.129</td>
<td>1</td>
<td>1.29</td>
<td>1.31</td>
<td>1.02</td>
<td></td>
<td></td>
<td></td>
<td>1.02</td>
<td>1.13</td>
<td>1.17</td>
</tr>
<tr>
<td>06.08.13 13:16</td>
<td>2.26</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.26</td>
<td>1.02</td>
<td>1.43</td>
</tr>
<tr>
<td>06.08.13 14:12</td>
<td>2.539</td>
<td>1</td>
<td>1.52</td>
<td>1.47</td>
<td>1.13</td>
<td></td>
<td></td>
<td></td>
<td>1.34</td>
<td>1.15</td>
<td>1.39</td>
</tr>
<tr>
<td>06.08.13 15:12</td>
<td>2.735</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.59</td>
<td>1.16</td>
<td>1.59</td>
</tr>
<tr>
<td>06.08.13 15:59</td>
<td>2.864</td>
<td>1</td>
<td>1.67</td>
<td>1.59</td>
<td>1.17</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>06.08.13 16:35</td>
<td>2.872</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.51</td>
<td>1.46</td>
<td>1.42</td>
</tr>
<tr>
<td>07.08.13 08:33</td>
<td>1.85</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.05</td>
<td>0.85</td>
<td>1.50</td>
</tr>
<tr>
<td>07.08.13 09:02</td>
<td>1.87</td>
<td>1</td>
<td>1.18</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.01</td>
<td>0.83</td>
<td>1.45</td>
</tr>
<tr>
<td>08.10.13 15:12</td>
<td>0.214</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.30</td>
<td>0.26</td>
<td>0.38</td>
</tr>
<tr>
<td>08.10.13 16:10</td>
<td>0.238</td>
<td>1</td>
<td>0.35</td>
<td>0.41</td>
<td>0.33</td>
<td></td>
<td></td>
<td></td>
<td>0.31</td>
<td>0.26</td>
<td></td>
</tr>
<tr>
<td>08.10.13 17:16</td>
<td>0.243</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.32</td>
<td>0.28</td>
<td></td>
</tr>
<tr>
<td>08.10.13 18:02</td>
<td>0.242</td>
<td>1</td>
<td>0.38</td>
<td>0.40</td>
<td>0.32</td>
<td></td>
<td></td>
<td></td>
<td>0.31</td>
<td>0.26</td>
<td></td>
</tr>
<tr>
<td>09.10.13 09:18</td>
<td>0.146</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.23</td>
<td>0.21</td>
<td>0.30</td>
</tr>
<tr>
<td>09.10.13 09:53</td>
<td>0.142</td>
<td>1</td>
<td>0.31</td>
<td>0.30</td>
<td>0.23</td>
<td></td>
<td></td>
<td></td>
<td>0.21</td>
<td>0.23</td>
<td>0.28</td>
</tr>
<tr>
<td>09.10.13 11:38</td>
<td>0.138</td>
<td>1</td>
<td>0.31</td>
<td>0.29</td>
<td>0.23</td>
<td>0.23</td>
<td>0.23</td>
<td>0.21</td>
<td>0.30</td>
<td>0.32</td>
<td>0.34</td>
</tr>
<tr>
<td>09.10.13 12:46</td>
<td>0.135</td>
<td>1</td>
<td>0.32</td>
<td>0.31</td>
<td>0.25</td>
<td>0.25</td>
<td>0.25</td>
<td>0.21</td>
<td>0.33</td>
<td>0.35</td>
<td>0.38</td>
</tr>
<tr>
<td>09.10.13 14:13</td>
<td>0.168</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.25</td>
<td>0.37</td>
<td>0.40</td>
</tr>
</tbody>
</table>
### Table S1 continued

<table>
<thead>
<tr>
<th>Date/Time</th>
<th>Q</th>
<th>Inj.</th>
<th>Flow Velocity (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>dd.mm.yy hh:mm</td>
<td>(m³/s) Loc</td>
<td>R#01</td>
<td>R#02</td>
</tr>
<tr>
<td>09.10.13 16:26</td>
<td>0.247</td>
<td>1</td>
<td>0.40</td>
</tr>
<tr>
<td>09.10.13 16:50</td>
<td>0.251</td>
<td>1</td>
<td>0.40</td>
</tr>
</tbody>
</table>

1) Dye tracer injection location 1 is 92 m upstream of R#01 on the glacier forefield, location 2 is 43 m upstream of R#06, above the knickpoint to the steep reach.

2) Empty cells refer to no measurements available, either due to injection locations that are downstream of the fluorometer, or due to fluorometers that did not function properly (e.g., filled/covered with sediments, displaced items due to water flow, invalid breakthrough curves, or battery and data logger problems).

### Table S2: Hydraulic geometry relations

<table>
<thead>
<tr>
<th>ID</th>
<th>k</th>
<th>m</th>
<th>r²</th>
<th>SE</th>
<th>a</th>
<th>b</th>
<th>r²</th>
<th>SE</th>
<th>w=( aQ^0 )</th>
<th>w_p=( aQ^0 )</th>
<th>d_h=( aQ^0 )</th>
<th>d=( aQ^0 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>R#01</td>
<td>0.95</td>
<td>0.58</td>
<td>0.97</td>
<td>0.13</td>
<td>5.54</td>
<td>0.11</td>
<td>0.93</td>
<td>0.19</td>
<td>5.63</td>
<td>0.12</td>
<td>0.93</td>
<td>0.20</td>
</tr>
<tr>
<td>R#02</td>
<td>0.86</td>
<td>0.54</td>
<td>1.00</td>
<td>0.02</td>
<td>6.95</td>
<td>0.25</td>
<td>1.00</td>
<td>0.08</td>
<td>7.10</td>
<td>0.25</td>
<td>1.00</td>
<td>0.08</td>
</tr>
<tr>
<td>R#03</td>
<td>0.67</td>
<td>0.51</td>
<td>0.99</td>
<td>0.03</td>
<td>7.36</td>
<td>0.18</td>
<td>0.98</td>
<td>0.18</td>
<td>7.64</td>
<td>0.19</td>
<td>0.97</td>
<td>0.24</td>
</tr>
<tr>
<td>R#04</td>
<td>0.69</td>
<td>0.54</td>
<td>1.00</td>
<td>0.02</td>
<td>6.95</td>
<td>0.17</td>
<td>0.99</td>
<td>0.10</td>
<td>7.23</td>
<td>0.18</td>
<td>0.99</td>
<td>0.12</td>
</tr>
<tr>
<td>R#05</td>
<td>0.73</td>
<td>0.55</td>
<td>1.00</td>
<td>0.02</td>
<td>6.53</td>
<td>0.16</td>
<td>0.99</td>
<td>0.06</td>
<td>6.82</td>
<td>0.16</td>
<td>1.00</td>
<td>0.06</td>
</tr>
<tr>
<td>R#06</td>
<td>0.70</td>
<td>0.59</td>
<td>0.99</td>
<td>0.08</td>
<td>5.55</td>
<td>0.17</td>
<td>0.97</td>
<td>0.20</td>
<td>5.85</td>
<td>0.18</td>
<td>0.97</td>
<td>0.22</td>
</tr>
<tr>
<td>R#07</td>
<td>0.70</td>
<td>0.61</td>
<td>0.97</td>
<td>0.10</td>
<td>6.11</td>
<td>0.15</td>
<td>0.93</td>
<td>0.28</td>
<td>6.29</td>
<td>0.14</td>
<td>0.92</td>
<td>0.28</td>
</tr>
<tr>
<td>R#08</td>
<td>0.88</td>
<td>0.54</td>
<td>0.98</td>
<td>0.08</td>
<td>3.79</td>
<td>0.29</td>
<td>0.97</td>
<td>0.20</td>
<td>4.67</td>
<td>0.28</td>
<td>0.97</td>
<td>0.23</td>
</tr>
<tr>
<td>R#09</td>
<td>0.86</td>
<td>0.49</td>
<td>0.99</td>
<td>0.04</td>
<td>4.16</td>
<td>0.29</td>
<td>0.99</td>
<td>0.12</td>
<td>4.91</td>
<td>0.29</td>
<td>0.99</td>
<td>0.14</td>
</tr>
<tr>
<td>R#10</td>
<td>0.85</td>
<td>0.44</td>
<td>0.98</td>
<td>0.06</td>
<td>5.63</td>
<td>0.26</td>
<td>0.95</td>
<td>0.32</td>
<td>5.90</td>
<td>0.26</td>
<td>0.95</td>
<td>0.35</td>
</tr>
</tbody>
</table>

Mean 0.80 0.54 0.99 0.06 5.67 0.22 0.95 0.21 6.05 0.22 0.95 0.24 0.24 0.25 0.96 0.01 0.22 0.25 0.96 0.01

STD 0.09 0.04 0.01 0.03 1.16 0.06 0.08 0.16 0.95 0.06 0.08 0.17 0.05 0.06 0.05 0.01 0.03 0.05 0.05 0.01

1) \( Q \)=flow discharge rate; \( v \)=velocity, \( w \)=width; \( w_p \)=wetted perimeter; \( d_h \)=flow depth; \( d \)=hydraulic radius. Note: Relations are reach-averaged.
SUPPLEMENTARY MATERIAL: STUDY II (CHAPTER III)

Applicability of bedload transport models for mixed size sediments in steep streams considering macro-roughness

Johannes M. Schneider\textsuperscript{1,2}, johannes.schneider@wsl.ch (corresponding author)
Dieter Rickenmann\textsuperscript{1}, dieter.rickenmann@wsl.ch
Jens M. Turowski\textsuperscript{1,3}, jens.turowski@wsl.ch
Kristin Bunte\textsuperscript{4}, kbunte@ engr.colostate.edu
James W. Kirchner\textsuperscript{1,2}, kirchner@ethz.ch

\textsuperscript{1) Swiss Federal Research Institute WSL, Mountain Hydrology and Mass Movements, Zürcherstrasse 111, 8903 Birmensdorf, Switzerland
\textsuperscript{2) Swiss Federal Institute of Technology ETH Zürich, Department of Environmental Systems Science, 8092 Zürich, Switzerland
\textsuperscript{3) Helmholtz Centre Potsdam, GFZ German Research Centre for Geosciences, Telegrafenberg 14473 Potsdam, Germany
\textsuperscript{4) Colorado State University, Engineering Research Center, 1320 Campus Delivery, Fort Collins, CO 80523, USA

The auxiliary material contains additional figures (Figures S1, S2, S6, S8, S9 and S10) and table S1 referring to the results derived from the Wilcock et al. [2009] approach instead from the Rickenmann and Recking [2011] approach to account for flow resistance and to reduce the total bed shear stress.

Figures S3, S4 and S5 show the hiding exponent $b$ related to bed characteristics when using the total bed shear stress (Fig. S3), the reduced bed shear stress of Rickenmann and Recking [2011] (Fig. S4) and the effective shear stress of Wilcock et al., [2009] (Fig. S5).

Figure S7 shows the similarity collapse of total dimensionless transport rates as suggested by Mueller et al. [2005], for the total shear stress and the effective shear stress (derived according to Rickenmann and Recking [2011]).

REFERENCES

### Table S1: a) Hiding function and dimensionless fractional bedload rating curves derived from the reduced effective shear stress according to Wilcock [2009] (WC2009). b) Dimensionless bedload rating curves derived from total transport rates \(\tau > 4\) mm and an reduced effective shear stress according to WC2009.

<table>
<thead>
<tr>
<th>Stream</th>
<th>(\tau^{*}<em>{r</em>{D50}})</th>
<th>(b)</th>
<th>(r^2)</th>
<th>(m_i)</th>
<th>(\tau^{*}<em>{r</em>{D50tot}})</th>
<th>(m_{tot})</th>
<th>(\tau^{*}<em>{r</em>{D50tot}})</th>
<th>(r^2)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Erlenbach</td>
<td>0.005</td>
<td>-0.84</td>
<td>0.928</td>
<td>2.54</td>
<td>0.006</td>
<td>2.74</td>
<td>0.026</td>
<td>6.90</td>
</tr>
<tr>
<td>Riedbach</td>
<td>0.029</td>
<td>-0.87</td>
<td>0.978</td>
<td>7.99</td>
<td>0.026</td>
<td>6.90</td>
<td>0.026</td>
<td>6.90</td>
</tr>
<tr>
<td>St. Louis Cr.</td>
<td>0.051</td>
<td>-0.95</td>
<td>0.993</td>
<td>3.67</td>
<td>0.045</td>
<td>6.43</td>
<td>0.026</td>
<td>18.77</td>
</tr>
<tr>
<td>Cherry Cr.</td>
<td>0.048</td>
<td>-0.92</td>
<td>0.998</td>
<td>11.19</td>
<td>0.045</td>
<td>13.10</td>
<td>0.026</td>
<td>18.77</td>
</tr>
<tr>
<td>Litl.Granite Cr.A</td>
<td>0.026</td>
<td>-0.95</td>
<td>1.000</td>
<td>14.22</td>
<td>0.026</td>
<td>18.77</td>
<td>0.026</td>
<td>18.77</td>
</tr>
<tr>
<td>E.St.Louis Cr.A</td>
<td>0.018</td>
<td>-0.95</td>
<td>0.993</td>
<td>3.67</td>
<td>0.045</td>
<td>6.43</td>
<td>0.026</td>
<td>18.77</td>
</tr>
<tr>
<td>E.St.Louis Cr.B</td>
<td>0.020</td>
<td>-0.97</td>
<td>0.999</td>
<td>8.56</td>
<td>0.021</td>
<td>9.67</td>
<td>0.021</td>
<td>9.67</td>
</tr>
<tr>
<td>Halfmoon Cr.</td>
<td>0.028</td>
<td>-0.91</td>
<td>0.997</td>
<td>4.78</td>
<td>0.026</td>
<td>7.70</td>
<td>0.026</td>
<td>7.70</td>
</tr>
<tr>
<td>Hayden Cr.</td>
<td>0.030</td>
<td>-0.90</td>
<td>0.998</td>
<td>7.46</td>
<td>0.030</td>
<td>8.27</td>
<td>0.030</td>
<td>8.27</td>
</tr>
<tr>
<td>E.Dallas Cr.</td>
<td>0.026</td>
<td>-0.70</td>
<td>0.997</td>
<td>6.96</td>
<td>0.020</td>
<td>6.23</td>
<td>0.020</td>
<td>6.23</td>
</tr>
<tr>
<td>Fool Cr.</td>
<td>0.012</td>
<td>-0.97</td>
<td>0.999</td>
<td>12.39</td>
<td>0.015</td>
<td>12.21</td>
<td>0.015</td>
<td>12.21</td>
</tr>
<tr>
<td>NF Swan Cr.</td>
<td>0.039</td>
<td>-1.16</td>
<td>0.984</td>
<td>2.48</td>
<td>0.057</td>
<td>3.94</td>
<td>0.057</td>
<td>3.94</td>
</tr>
<tr>
<td>Litl.Granite Cr.B</td>
<td>0.022</td>
<td>-0.82</td>
<td>0.995</td>
<td>6.23</td>
<td>0.020</td>
<td>8.10</td>
<td>0.020</td>
<td>8.10</td>
</tr>
<tr>
<td>Oak Cr.</td>
<td>0.021</td>
<td>-0.73</td>
<td>0.999</td>
<td>5.34</td>
<td>0.018</td>
<td>5.12</td>
<td>0.018</td>
<td>5.12</td>
</tr>
<tr>
<td>Big W.R..nr.Ket.</td>
<td>0.014</td>
<td>-0.86</td>
<td>0.987</td>
<td>3.27</td>
<td>0.019</td>
<td>4.76</td>
<td>0.019</td>
<td>4.76</td>
</tr>
<tr>
<td>Boise R.</td>
<td>0.010</td>
<td>-0.86</td>
<td>0.986</td>
<td>2.85</td>
<td>0.033</td>
<td>3.90</td>
<td>0.033</td>
<td>3.90</td>
</tr>
<tr>
<td>Dollar Cr.</td>
<td>0.015</td>
<td>-0.81</td>
<td>0.977</td>
<td>1.77</td>
<td>0.041</td>
<td>1.93</td>
<td>0.041</td>
<td>1.93</td>
</tr>
<tr>
<td>Herd Cr.</td>
<td>0.021</td>
<td>-0.91</td>
<td>0.996</td>
<td>1.56</td>
<td>0.010</td>
<td>2.69</td>
<td>0.010</td>
<td>2.69</td>
</tr>
<tr>
<td>Lochsa R.</td>
<td>0.019</td>
<td>-0.79</td>
<td>0.991</td>
<td>4.88</td>
<td>0.020</td>
<td>6.25</td>
<td>0.020</td>
<td>6.25</td>
</tr>
<tr>
<td>MF Red R.</td>
<td>0.029</td>
<td>-0.06</td>
<td>0.023</td>
<td>1.67</td>
<td>0.037</td>
<td>1.01</td>
<td>0.037</td>
<td>1.01</td>
</tr>
<tr>
<td>MF Salmon R.</td>
<td>0.018</td>
<td>-0.91</td>
<td>0.994</td>
<td>8.80</td>
<td>0.020</td>
<td>12.24</td>
<td>0.020</td>
<td>12.24</td>
</tr>
<tr>
<td>NFCLI.Water R.</td>
<td>0.006</td>
<td>-1.08</td>
<td>0.993</td>
<td>4.84</td>
<td>0.011</td>
<td>7.15</td>
<td>0.011</td>
<td>7.15</td>
</tr>
<tr>
<td>Rapid R.</td>
<td>0.025</td>
<td>-0.44</td>
<td>0.842</td>
<td>1.86</td>
<td>0.025</td>
<td>2.52</td>
<td>0.025</td>
<td>2.52</td>
</tr>
<tr>
<td>Salmon Rbl.Ynk.F</td>
<td>0.015</td>
<td>-0.96</td>
<td>0.985</td>
<td>5.90</td>
<td>0.022</td>
<td>8.53</td>
<td>0.022</td>
<td>8.53</td>
</tr>
<tr>
<td>Salmon R. nr. Obs.</td>
<td>0.021</td>
<td>-0.82</td>
<td>0.992</td>
<td>4.97</td>
<td>0.025</td>
<td>7.52</td>
<td>0.025</td>
<td>7.52</td>
</tr>
<tr>
<td>Salmon R. nr. Shp.</td>
<td>0.012</td>
<td>-0.93</td>
<td>0.997</td>
<td>6.74</td>
<td>0.016</td>
<td>8.63</td>
<td>0.016</td>
<td>8.63</td>
</tr>
<tr>
<td>Selway R.</td>
<td>0.013</td>
<td>-0.92</td>
<td>0.994</td>
<td>4.99</td>
<td>0.015</td>
<td>9.47</td>
<td>0.015</td>
<td>9.47</td>
</tr>
<tr>
<td>SF Payette R.</td>
<td>0.005</td>
<td>-0.81</td>
<td>0.704</td>
<td>1.11</td>
<td>0.018</td>
<td>2.54</td>
<td>0.018</td>
<td>2.54</td>
</tr>
<tr>
<td>Squaw Cr.USGS</td>
<td>0.017</td>
<td>-0.74</td>
<td>0.927</td>
<td>2.43</td>
<td>0.025</td>
<td>4.35</td>
<td>0.025</td>
<td>4.35</td>
</tr>
<tr>
<td>Thompson Cr.</td>
<td>0.022</td>
<td>-0.62</td>
<td>0.950</td>
<td>3.02</td>
<td>0.025</td>
<td>2.92</td>
<td>0.025</td>
<td>2.92</td>
</tr>
<tr>
<td>Susitna R.nr.Talk..</td>
<td>0.011</td>
<td>-1.03</td>
<td>0.876</td>
<td>2.46</td>
<td>0.024</td>
<td>5.94</td>
<td>0.024</td>
<td>5.94</td>
</tr>
<tr>
<td>Talkeetna R.nr.Talk..</td>
<td>0.006</td>
<td>-0.72</td>
<td>0.967</td>
<td>3.66</td>
<td>0.008</td>
<td>4.85</td>
<td>0.008</td>
<td>4.85</td>
</tr>
<tr>
<td>Lo. NF Cabin Cr.</td>
<td>0.044</td>
<td>-0.23</td>
<td>0.252</td>
<td>0.33</td>
<td>0.098</td>
<td>0.60</td>
<td>0.098</td>
<td>0.60</td>
</tr>
<tr>
<td>Buffalo Cr.</td>
<td>0.009</td>
<td>-0.89</td>
<td>0.995</td>
<td>5.11</td>
<td>0.009</td>
<td>5.21</td>
<td>0.009</td>
<td>5.21</td>
</tr>
<tr>
<td>Horse Cr.</td>
<td>0.007</td>
<td>-0.39</td>
<td>0.853</td>
<td>3.19</td>
<td>0.004</td>
<td>2.59</td>
<td>0.004</td>
<td>2.59</td>
</tr>
<tr>
<td>Lo. SF Williams F.</td>
<td>0.011</td>
<td>-0.42</td>
<td>0.351</td>
<td>1.57</td>
<td>0.017</td>
<td>2.33</td>
<td>0.017</td>
<td>2.33</td>
</tr>
<tr>
<td>Williams F.</td>
<td>0.006</td>
<td>-0.68</td>
<td>0.938</td>
<td>3.33</td>
<td>0.008</td>
<td>4.75</td>
<td>0.008</td>
<td>4.75</td>
</tr>
</tbody>
</table>

| Median Main-Dataset     | 0.026                   | -0.91 | 6.96   | 0.023 | 7.9                    |
| Median HS-Dataset       | 0.014                   | -0.74 | 3.19   | 0.02  | 4.75                   |

**Appendix**
Figure S1: Dimensionless reference shear stress ($\tau^*_r$) related to stream characteristics. $\tau^*_r$ derived from the reduced effective acting bed shear stress $\tau^*_b$ [Wilcock et al., 2009] and fractional transport rates. a) channel bed slope, with the empirical relations of Bunte et al. [2013], Lamb et al. [2008]; and Mueller et al. [2005]; b) Sand content with the empirical relation of Wilcock and Crowe [2003] (WC2003); c) $D_{94\text{surf}}$ of the surface grain size distribution (GSD); d) $D_{50\text{surf}}/D_{50\text{sub}}$ as a measure of the degree of armouring with the empirical relation of Efthymiou [2012]; e) $D_{84\text{surf}}/D_{30\text{surf}}$ as a measure of the width of the bed surface GSD; f) Geometric standard deviation of the bed surface GSD with the empirical relation of Gaeuman et al. [2009].
Figure S2: Dimensionless reference shear stress ($\tau^*_{rD50}$) related to channel bed slope for the reduced effective shear stress $\tau^*_{D50}$ according to Wilcock et al., 2009 (WC2009). Filled circles correspond to $\tau^*_{D50}$ values from analysing fractional transport rates (see reference approach Section (2.3.2) and Figure (2b)). Crosses indicate $\tau^*_{D50}$ derived from analysis of total dimensionless transport rates. Black line was fitted to the Main Dataset, and the gray line to the HS Dataset. Dashed blue line in b) corresponds to the median $\tau^*_{D50}$ from the Main Dataset.
Figure S3: Hiding exponent $|b|$ related to stream characteristics based on the total boundary shear stress $\tau^*$. a) channel bed slope; b) Sand content; c) $D_{84\text{surf}}$ of the surface grain-size distribution (GSD); d) $D_{50\text{surf}}/D_{50\text{sub}}$ as a measure of the degree of armouring; e) $D_{84\text{surf}}/D_{30\text{surf}}$ as a measure of the width of the bed surface GSD; f) Geometric standard deviation of the bed surface GSD.
Figure S4: Hiding exponent $|b|$ related to stream characteristics based on the reduced effective shear stress $\tau^*$ according to Rickenmann and Recking [2011]. a) channel bed slope; b) Sand content; c) $D_{84\text{Surf}}$ of the surface grain-size distribution (GSD); d) $D_{50\text{Surf}}/D_{50\text{Sub}}$ as a measure of the degree of armouring; e) $D_{84\text{Surf}}/D_{30\text{Surf}}$ as a measure of the width of the bed surface GSD; f) Geometric standard deviation of the bed surface GSD.
Figure S5: Hiding exponent $|b|$ related to stream characteristics based on the reduced effective shear stress $\tau^*$ according to Wilcock et al. [2009].

a) Channel bed slope; b) Sand content; c) $D_{84Surf}$ of surface grain-size distribution (GSD); d) $D_{50Surf}/D_{50Sub}$ as a measure of the degree of armouring; e) $D_{84Surf}/D_{84Sub}$ as a measure of the width of the bed surface GSD; f) Geometric standard deviation of the bed surface GSD.
Figure S6: Similarity collapse of fractional transport rates based on $\tau^*_{ri}$ (derived from the reference approach of section (2.3.2), step (ii)) and using the reduced shear stress from Wilcock et al. [2009].
Figure S7: Similarity collapse of total dimensionless transport rates based on $\tau^*_{rD50}$ (derived according to Mueller et al. [2005], see section (2.3.2)). a) Collapse based on the total boundary shear stress $\tau^*$. b) Collapse based on the reduced effective bed shear stress $\tau^{*'}$ (RR2011) [Rickenmann and Recking, 2011].
Figure S8: Similarity collapse of total dimensionless transport rates based on $\tau^{*}_{rD50}$ (derived according to Mueller et al. [2005], see section (2.3.2)) and using the reduced shear stress $\tau^{*'}$ from Wilcock et al. [2009].
Figure S9: Bedload transport predictions (> 4 mm) based on the reduced effective acting bed shear stress \( \tau^* \) [Wilcock et al., 2009] for a) the Wilcock and Crowe [2003] equation; b) a modified fractional transport equation where \( \phi_i = \tau^* / \tau^*_{rD50} \) in combination with a constant reference shear stress \( \tau^*_{rD50} \); c) a total transport equation \( \phi = \tau^*_{D50} / \tau^* \) in combination with a constant \( \tau^*_{D50} \).
Figure S10: Discrepancy ratios (calculated relative to measured transport rates) related to channel slope. Calculations based on the reduced effective acting bed shear stress $\tau^*_{\text{eff}}$ for a) the Wilcock and Crowe [2003] equation; b) a modified fractional transport equation where $\phi_i = \tau^*_i / \tau^*_{\text{ref}}$ in combination with a constant reference shear stress $\tau^*_r D_{50}$; c) a total transport equation $\phi = \tau^*_{D_{50}} / \tau^*_{r D_{50}}$ in combination with a constant $\tau^*_{r D_{50}}$. 
SUPPLEMENTARY MATERIAL: STUDY III (CHAPTER IV)

Scaling relationships between bedload volumes, transport distances and stream power in steep mountain channels

Johannes M. Schneider¹,², johannes.schneider@wsl.ch
Jens M. Turowski³,¹, jens.turowski@gfz-potsdam.de
Dieter Rickenmann¹, dieter.rickenmann@wsl.ch
Ramon Hegglin¹, ramonhegglin@gmx.ch
Sabrina Arrigo¹, sabrina.arrigo@nipo.ch
Luca Mao⁴, lmao@uc.cl
James W. Kirchner²,¹, kirchner@ethz.ch

1) Swiss Federal Research Institute WSL, Mountain Hydrology and Mass Movements, Zürcherstrasse 111, 8903 Birmensdorf, Switzerland
2) Swiss Federal Institute of Technology ETH Zürich, Department of Environmental System Science, 8092 Zürich, Switzerland
3) Helmholtz Centre Potsdam, GFZ German Research Centre for Geosciences, Telegrafenberg 14473 Potsdam, Germany
4) Pontificia Universidad Católica de Chile, Department of Ecosystems and Environment, Av. Vicuña Mackenna 4860, Macul, Santiago, Chile

S1. Tracer particle preparation (Erlenbach only)
(Complementing Section 2.3.1.1)

For the Erlenbach, a total of 540 artificial and natural tracer particles were prepared in five classes with mean weights of: 0.1kg +/- 5% (rods); 0.1kg +/-5% (discs); 0.5kg +/- 62%; 1.5kg +/- 36% and 3.4kg +/- 26%. The 0.1 kg discs and rods were made of synthetic resin with a density of 2.6 g/cm³, whereas all other particles were prepared from natural stones. In addition, all stones were marked with a number and painted to help find them more easily in the channel bed surface.

A total of 760 natural particles in five size classes were marked with RFID transponders and painted. The smallest particle size prepared with an RFID transponder was 28 mm in b-axis. The size classes included mean weights of 0.05kg, 0.1kg, 0.2 kg, 0.4kg, 0.8kg and 1.6kg, with the weight of individual particles varying by +/-10% around the mean value. Approximately the same number of tracer particles (~150) was prepared in each weight class. The stones were simply placed on the surface of the streambed.
S2. Tracer recovery rates in the Erlenbach  
(Complementing Section 2.3.1.2)

The low Erlenbach recovery rates compared to studies previously performed in gravel-bed rivers and mountain streams [e.g., Lamarre and Roy, 2008; Lamarre et al., 2005; Liébault et al., 2012] might be explained by the Erlenbach’s complex bed topography and high-energy flow environment.

The bed of the Erlenbach features many large boulders and abundant woody debris, making it difficult to access the entire streambed and to maneuver the RFID antenna (0.8 m in diameter). The channel slope and the grain size distribution of the Rio Cordon (RC) are comparable to those of the Erlenbach, but the RC channel is wider and has significantly less woody debris, which could explain the generally better recovery rates in the RC studies.

The low recovery rates at the Erlenbach can also be attributed to the large flood events observed during the 2-year RFID campaign. Four events occurred with peak discharges \( Q_p > 3 \) m\(^3\)/s (corresponding to a recurrence interval of \( \sim 3 \) years) and one event occurred with a peak discharge of \( \sim 10 \) m\(^3\)/s (corresponding to a recurrence interval of \( \sim 20 \) years). It is likely that many particles were missed in the searches because they were flushed out of the study reach during these large flood events. This assumption is supported by the high loss of tracer particles during these events (for example Event 1 and 2, table 2). It is also likely that many particles were buried at a depth that was beyond of the detection range of the RFID antenna (about 0.6 m). For example, based on consecutive long-profile measurements Turowski et al. [2013] report deposition and erosion depths locally exceeding 2 m for the 10 m\(^3\)/s event in August 2010. The assumption that particles were buried or otherwise deposited out of the detection range of the antenna is also supported by the observed re-emergence rates: up to 35% (min 0%; mean 11%) of tracer particles that were not recovered in one survey re-appeared in a later survey.

A last reason for occasional low recovery rates, not related to steep mountain stream characteristics, is "multi-tag collision": when several RFID tags are close together within the antenna field, the individual signals interfere with each other, preventing clear identification of individual tracers. For example, during Event 8 with a very low flow magnitude, only a few tracers moved from their initial position (tracers where placed before this event at two spots) making it impossible to identify each individual particle.

S3. Estimating the transport distances of unmeasured grain sizes  
(Complementing Section 2.3.1.3)

Because transport distances increase with decreasing particle size [cf. Church and Hassan, 1992; Haschenburger, 1996; Haschenburger and Church, 1998; Hassan et al., 1992; Lenzi, 2004; Wilcock, 1997], two main uncertainties arise when tracer data are used to infer the mean transport distance of the total bedload:

1. The collected tracer grain sizes are not always representative for the grain-size distribution of the tracer population itself, i.e. the mean b-axis of the collected tracers may not always correspond to the mean b-axis of the tracer population. Depending on the tracer grain sizes detected in each event, this might distort the estimated mean transport distances because
Study III

the smaller grains travel farther.

(2) The tracer grain-size distribution (GSD), ranging from 28 to 160 mm, is narrower than the average bedload GSD, which ranges from ~10 to 300 mm (Figure 2). Thus the observed transport distances of the tracer particles are only representative of roughly 60% of the total bedload.

In short: the grain size distribution of the observed tracer population changes from transport event to transport event, and it is generally coarser than the average bedload grain size distribution. Therefore, the derived mean transport distances of the tracer particles do not correspond to the mean transport distances of the total bedload.

To estimate transport distances of the grain size fractions that were not represented by the measured tracer particles ($L_{\text{Obs}}$), and thus to estimate the transport distances of the total bedload ($L_B$), we used the empirical relation of Church and Hassan [1992] (Equation S1), which relates tracer transport distances to particle grain size within the range $0.1 < D_i/D_{\text{50Sub}} < 10$. The Church and Hassan [1992] equation is already stated in main text (equation 1), but we repeat the definition here because of its importance for the following analysis. The equation scales the transport distances ($L^*$) of the tracer particles to scaled grain sizes.

$$L^* = L_i / L_{D_{\text{50Surf}}} = 1.77 \left[ 1 - \log_{10} \left( D_i / D_{\text{50Surf}} \right) \right]^{1.35}$$  \hspace{1cm} (S1)

Where $L_i$ is the transport distance of the individual grain with the diameter $D_i$; $L_{D_{\text{50Surf}}}$ is the mean transport distance of the grain size fraction containing the $D_{\text{50Surf}}$ and can be seen as a measure of flood event size [Haschenburger, 1996].

Firstly, using Equation S1 (Equation 3 in the paper) we determined scaled transport distances $L^*_{\text{Obs}}$ for the grain size distribution of the observed mean tracer transport distances ($L_{\text{Obs}}$) with a median grain size $D_i = D_{\text{50Obs}}$ for each transport event ($D_{\text{50Obs}} =$median grain size of observed tracer population). In addition we determined from Equation S1 scaled transport distances $L^*_{\text{T}}$ for the true grain size distribution of the complete tracer population using a $D_i = D_{\text{50T}}$ ($D_{\text{50T}} =$median grain size of complete tracer population). To estimate the transport distances of the complete tracer population ($L_T$, Equation S2a) we scaled mean observed transport $L_{\text{Obs}}$ distances with the $L^*_{\text{T}}/L^*_{\text{Obs}}$ ratio.

$$L_T = L_{\text{T}}^* \frac{L_{\text{Obs}}^*}{L_{\text{Obs}}} ; L_F = L_{\text{F}}^* \frac{L_{\text{Obs}}^*}{L_{\text{Obs}}} ; L_C = L_{\text{C}}^* \frac{L_{\text{Obs}}^*}{L_{\text{Obs}}}$$  \hspace{1cm} (S2a, b, c)

Secondly, and using the same procedure as in the first step, for each transport event we estimated the mean transport distances for the finer fractions ($L_F$, Equation S2b) with $D_i = D_{\text{50F}}$ ($D_{\text{50F}} =$ median grain size of the fine fraction, $D = 10$-28 mm) and (c) the coarser fractions ($L_C$, Equation S2c) with $D_i = D_{\text{50C}}$($D_{\text{50C}} =$ median grain size of the coarse fraction, $D> 160$ mm), not measured by the tracer particles.

The fractions and their relative proportions, according to the GSD of the sediment collected in the retention basin (see also Figure 2 in the paper), are: (i) fine material, $D = 10$-28 mm, comprising 30% of the total, (ii) the fraction covered by the tracer population, $D = 28$-160 mm, comprising 60% of the total, and (iii) the coarse fraction, $D> 160$ mm, comprising 10% of the total. We use these percentages in the following relationship to
estimate $L_B$:

$$L_B = 0.3L_F + 0.6L_T + 0.1L_C$$ (S3)

The use of only three size fractions might appear to be a crude approximation, however in view of the other uncertainties (in measured transport distances, for example, or in Equation S1 itself) we assume that a finer resolution of the grain size fractions would not improve accuracy.

S4. Measured bedload volumes by the Erlenbach PBIS/Geophone system
(Complementing Section 2.4.1)

The PBIS/Geophone system is a well-established indirect method that has been used to measure bedload transport in several streams [e.g., Turowski and Rickenmann, 2011; Turowski et al., 2011].

The sensors are installed in an array of steel plates over the entire cross-section of a check-dam immediately above the sediment retention basin. The sensors register vibrations produced by impacting stones, and the number of peaks above a pre-defined threshold is stored for each minute when transport occurs. Only grains with diameters greater than about 20 mm produce detectable vibrations that can be clearly distinguished from background noise. For calibration to absolute transport rates, the PBIS/Geophone impulses were compared with long-term records of sediment volumes in the sediment retention basin [Rickenmann and McArdell, 2007] and with short-term measurements of sediment volumes obtained from a moving bedload basket system [Rickenmann et al., 2012]. The recorded numbers of impulses are proportional to the recorded sediment volumes, allowing sediment transport loads to be inferred from the PBIS/Geophone system. The accuracy of the system allows sediment loads of entire flood events to be estimated within a factor of about two [Rickenmann and McArdell, 2007; 2008; Rickenmann et al., 2012].

Because the geophone data are calibrated against sediment volumes accumulated in the retention basin (including fine material and pore volume), estimating bedload transport requires subtracting the percentage of fine material transported in suspension rather than as bedload. The average percentage of suspended load (particle sizes smaller than 10 mm) was estimated as 42% of the total sediment transported into the sediment retention basin (Figure 2). The critical grain-size of 10 mm was derived by comparing the settling velocity $w_s$ [Wu and Wang, 2006] to the average shear velocity $U^*$ using a threshold value of $w_s/U^*$=0.8 [Dade and Friend, 1998] for the flow conditions when bedload transport begins. This determination of the critical grain-size, and the use of the average grain-size distribution collected in the retention basin, both involve significant uncertainties. As flow increases, the critical grain size should coarsen, as should the GSD of the total sediment transported. For example, using the same analysis for the maximum flow during the observation period (10 m$^3$/s) we estimate that a grain size of 24 mm (corresponding to the ~60$^{\text{th}}$ percentile of the grain size distribution) having been transported as suspended sediment. Furthermore, the estimated critical grain-size derived from $w_s/U^*$=0.8 is only a rough estimate, in view of the broad transition range from suspended to bedload as reported by Dade and Friend [1998] with 0.3<$w_s/U^*$ < 3. However, data limitations regarding the GSD of the transported sediment
per flood event prevent us from estimating the bedload fraction more accurately as a function of discharge.

S5 Processing of literature data
(Complementing Sections 3 and 4)

1. Lainbach, Germany [Gintz et al., 1996]
The Lainbach is a mountain stream with a channel slope of 0.02 m/m, draining a basin area of 15.2 km$^2$. Unit excess stream power and mean transport distances (including particles that did not move) were derived by manually exporting the data points from figure 5a [Gintz et al., 1996]. Derived unit excess stream power values were assigned to individual transport events according to the mean transport distances given in table 1. Given mean transport distances (of moved particles only) were compared with assigned unit excess stream power values.

2. Carnation Creek, Canada [Haschenburger, 2013; Haschenburger and Church, 1998]
Carnation Creek is a gravel-bed stream that drains a basin area of 11 km$^2$. Its channel slope ranges from 0.006-0.012 m/m at the study reach.

Total peak stream power (where total means power averaged over the stream width) as given in Haschenburger and Church [1998] was converted in unit excess stream power through division by given average active stream width (5.4 m) and subtraction of the critical unit stream power. Critical unit stream power was derived from the critical discharge of 6.7 m$^3$/s [Haschenburger 2013]. Cumulative unit excess stream energy values were derived by exporting cumulative total excess stream power (given in Haschenburger [2013], figure 5a) manually, dividing cumulative total excess stream power by average active stream width (5.4 m, Haschenburger and Church [1998]) and converting to J/m$^2$.

3. Ardennian Rivers [Houbrechts et al., 2012]
The Ardennian river dataset includes 15 streams (measured transport distances are available for four streams) with channel slopes ranging from 0.001 to 0.011 m/m and catchment sizes ranging from 12-2910 km$^2$.

Given unit total stream power was reduced by subtracting the critical stream power. Critical stream power was estimated from a critical dimensionless shear stress of 0.03 [-] which was converted to a critical unit discharge for each stream using the variable power (VPE) of Ferguson [2007]. Resulting negative unit excess stream power values have been ignored in the analysis.

REFERENCES

Appendix


### Table S1: Literature data

<table>
<thead>
<tr>
<th>Date</th>
<th>Stream/Reach</th>
<th>(Q_p) [(\text{m}^3/\text{s})]</th>
<th>(\omega_p) [(\text{W/m}^2)]</th>
<th>(\Sigma(\omega-\omega_c)) [(\text{J/m}^2)]</th>
<th>(L_B) [m]</th>
<th>(F_B) [(\text{m}^3)]</th>
<th>(f_B) [(\text{m}^3/\text{m})]</th>
<th>(h_B) [m]</th>
</tr>
</thead>
<tbody>
<tr>
<td>03.08.1988</td>
<td></td>
<td>4.3</td>
<td>42.5</td>
<td>15</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>13.08.1988</td>
<td></td>
<td>8.9</td>
<td>192.9</td>
<td>38</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>12.07.1989</td>
<td></td>
<td>8.3</td>
<td>173.3</td>
<td>60</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>14.07.1989</td>
<td></td>
<td>12.2</td>
<td>300.8</td>
<td>86</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>18.07.1989</td>
<td></td>
<td>3.4</td>
<td>13.1</td>
<td>7</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>28.07.1989</td>
<td></td>
<td>3.9</td>
<td>29.4</td>
<td>4</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>09.08.1989</td>
<td></td>
<td>3.4</td>
<td>13.1</td>
<td>9</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>28.06.1991</td>
<td></td>
<td>8.7</td>
<td>186.4</td>
<td>451</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>09.07.1991</td>
<td></td>
<td>3.6</td>
<td>19.6</td>
<td>16</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>25.07.1991</td>
<td>Reach 1</td>
<td>7.4</td>
<td>143.9</td>
<td>222</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>22.07.1992</td>
<td>Reach 1</td>
<td>10.2</td>
<td>235.4</td>
<td>271</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>20.10.1992</td>
<td>Reach 1</td>
<td>17.7</td>
<td>273.3</td>
<td>82.6</td>
<td>0.064</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>24.01.1993</td>
<td>Reach 1</td>
<td>36.3</td>
<td>465.0</td>
<td>125.9</td>
<td>0.24</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>04.03.1993</td>
<td>Reach 1</td>
<td>17.9</td>
<td>289.4</td>
<td>48.7</td>
<td>0.16</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>20.10.1992</td>
<td>Reach 2</td>
<td>17.7</td>
<td>300.2</td>
<td>49.3</td>
<td>0.074</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>24.01.1993</td>
<td>Reach 2</td>
<td>36.3</td>
<td>551.4</td>
<td>43.7</td>
<td>0.17</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>04.03.1993</td>
<td>Reach 2</td>
<td>17.9</td>
<td>244.0</td>
<td>70.3</td>
<td>0.077</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>29.08.1991</td>
<td>Reach 3</td>
<td>24.5</td>
<td>202.5</td>
<td>129.1</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>19.01.1991</td>
<td>Reach 3</td>
<td>30.4</td>
<td>255.8</td>
<td>98.5</td>
<td>0.078</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>29.01.1992</td>
<td>Reach 3</td>
<td>22.6</td>
<td>324.4</td>
<td>58.4</td>
<td>0.063</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>20.10.1992</td>
<td>Reach 3</td>
<td>17.7</td>
<td>128.2</td>
<td>26.8</td>
<td>0.053</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>24.01.1993</td>
<td>Reach 3</td>
<td>36.3</td>
<td>273.2</td>
<td>69.1</td>
<td>0.084</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>04.03.1993</td>
<td>Reach 3</td>
<td>17.9</td>
<td>142.8</td>
<td>25.8</td>
<td>0.047</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

a) Peak discharge (\(Q_p\)); Unit excess stream power of peak discharge (\(\omega_p\)); Cumulative unit excess stream energy (\(\Sigma(\omega-\omega_c)\)); Mean transport distances (\(L_B\)); Bedload volume (\(F_B\)); Measured active layer depth (\(h_B\)).
Appendix

Table S1: Literature data

<table>
<thead>
<tr>
<th>Date</th>
<th>Stream/Reach</th>
<th>(Q_p) [m(^3)/s]</th>
<th>(\omega_p - \omega_c) [W/m(^3)]</th>
<th>(\sum(\omega - \omega_c)) [J/m(^3)]</th>
<th>(L_B) [m]</th>
<th>(F_B) [m(^3)/m]</th>
<th>(f_B)</th>
<th>(h_B) [m]</th>
</tr>
</thead>
<tbody>
<tr>
<td>18.03.2005</td>
<td>Rulles</td>
<td>1.05</td>
<td>9.6</td>
<td></td>
<td>0.01</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>18.03.2008</td>
<td>Süre</td>
<td>1.4</td>
<td>-22.4</td>
<td></td>
<td>0.015</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>22.08.2007</td>
<td>Chavanne in VC</td>
<td>2.7</td>
<td>21.5</td>
<td>4.9</td>
<td>0.13</td>
<td>0.02</td>
<td>0.009</td>
<td></td>
</tr>
<tr>
<td>18.03.2005</td>
<td>Braunlauf</td>
<td>3.5</td>
<td>31.9</td>
<td></td>
<td>0.02</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>30.03.2008</td>
<td>E Ourthe in H.</td>
<td>16.8</td>
<td>1.8</td>
<td>29.9</td>
<td>4.8</td>
<td>0.61</td>
<td>0.027</td>
<td></td>
</tr>
<tr>
<td>19.01.2007</td>
<td>Berwinne in D.</td>
<td>18.9</td>
<td>26.3</td>
<td></td>
<td></td>
<td>0.037</td>
<td></td>
<td></td>
</tr>
<tr>
<td>17.02.2007</td>
<td>Berwinne in B.</td>
<td>19.1</td>
<td>63.3</td>
<td>17.9</td>
<td>1.6</td>
<td>0.23</td>
<td>0.03</td>
<td></td>
</tr>
<tr>
<td>17.01.2010</td>
<td>Berwinne in B.</td>
<td>11.4</td>
<td>15.3</td>
<td>11.8</td>
<td>0.3</td>
<td>0.04</td>
<td>0.015</td>
<td></td>
</tr>
<tr>
<td>16.08.2010</td>
<td>Berwinne in B.</td>
<td>6.6</td>
<td>-5.7</td>
<td>4.2</td>
<td>0.2</td>
<td>0.02</td>
<td>0.02</td>
<td></td>
</tr>
<tr>
<td>13.11.2010</td>
<td>Berwinne in B.</td>
<td>30.9</td>
<td>104.3</td>
<td>15.5</td>
<td>7.8</td>
<td>0.99</td>
<td>0.05</td>
<td></td>
</tr>
<tr>
<td>21.01.2005</td>
<td>Our</td>
<td>20</td>
<td>50.8</td>
<td></td>
<td></td>
<td>0.037</td>
<td></td>
<td></td>
</tr>
<tr>
<td>18.01.2007</td>
<td>Our</td>
<td>39.5</td>
<td>63.0</td>
<td></td>
<td></td>
<td>0.03</td>
<td></td>
<td></td>
</tr>
<tr>
<td>19.01.2007</td>
<td>Bolland</td>
<td>6.7</td>
<td>73.2</td>
<td></td>
<td></td>
<td>0.01</td>
<td></td>
<td></td>
</tr>
<tr>
<td>17.02.2009</td>
<td>Aizne Riv. Juz.</td>
<td>26.1</td>
<td>20.5</td>
<td>41.6</td>
<td>2.5</td>
<td>0.18</td>
<td>0.02</td>
<td></td>
</tr>
<tr>
<td>12.05.2009</td>
<td>Aizne Riv. Juz.</td>
<td>20.7</td>
<td>1.5</td>
<td>8.8</td>
<td>0.2</td>
<td>0.015</td>
<td>0.01</td>
<td></td>
</tr>
<tr>
<td>23.02.2010</td>
<td>Aizne Riv. Juz.</td>
<td>17</td>
<td>-10.5</td>
<td>13</td>
<td>1.4</td>
<td>0.10</td>
<td>0.025</td>
<td></td>
</tr>
<tr>
<td>09.01.2011</td>
<td>Aizne Riv. Juz.</td>
<td>48</td>
<td>93.5</td>
<td>297</td>
<td>241</td>
<td>17.64</td>
<td>0.06</td>
<td></td>
</tr>
<tr>
<td>18.01.2007</td>
<td>Amblève</td>
<td>67</td>
<td>72.5</td>
<td></td>
<td></td>
<td>0.037</td>
<td></td>
<td></td>
</tr>
<tr>
<td>18.01.2007</td>
<td>Salm</td>
<td>23</td>
<td>135.3</td>
<td></td>
<td></td>
<td>0.074</td>
<td></td>
<td></td>
</tr>
<tr>
<td>22.08.2007</td>
<td>Lesse</td>
<td>153.5</td>
<td>118.2</td>
<td></td>
<td></td>
<td>0.063</td>
<td></td>
<td></td>
</tr>
<tr>
<td>18.01.2007</td>
<td>Wayai</td>
<td>21.2</td>
<td>141.4</td>
<td></td>
<td></td>
<td>0.04</td>
<td></td>
<td></td>
</tr>
<tr>
<td>18.01.2007</td>
<td>Hoegne</td>
<td>52.7</td>
<td>135.9</td>
<td></td>
<td></td>
<td>0.08</td>
<td></td>
<td></td>
</tr>
<tr>
<td>24.02.2010</td>
<td>Lembrée</td>
<td>4.8</td>
<td>17.7</td>
<td></td>
<td></td>
<td>0.02</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

\[\]\(^{a}\) Peak discharge (\(Q_p\)); Unit excess stream power of peak discharge (\(\omega_p - \omega_c\)); Cumulative unit excess stream energy (\(\sum(\omega - \omega_c)\)); Mean transport distances (\(L_B\)); Bedload volume (\(F_B\)); Measured active layer depth (\(h_B\)).
PUBLICATION LIST

Thesis papers:

Schneider, J. M., D. Rickenmann, J. M. Turowski, and J. W. Kirchner (subm), Stream bed roughness and flow velocity in a steep mountain channel, submitted to Water Resources Research.


Additional Publications:


DANK

Mein Dank für die hilfreiche Unterstützung bei der Erstellung meiner Doktorarbeit geht zuallererst an meine Betreuer Dieter Rickenmann und Jens M. Turowski sowie meinen Doktorvater James W. Kirchner. Vielen Dank für die unzähligen Diskussionen, eure Ideen und Kritiken, die Überarbeitung meiner Manuskripte, eure Geduld und vor allem den Rückhalt, den ich während meiner Arbeit von Euch erfahren habe. Thanks also to Stuart Lane for the review of my thesis.

Mein weiterer Dank geht an Manfred Stähli und Alexandre Badoux für die angenehmen Rahmenbedingungen an der WSL, für die Ihr während meiner Doktorarbeit stets gesorgt habt sowie an die komplette FE Gebirgshydrologie und Massenbewegungen für das super Arbeitsklima. Vielen Dank an Bruno Fritschi, Pat Thee und Kari Steiner für den unermüdlichen technischen Support und die Hilfe bei den Feldarbeiten. Vielen Dank für die wertvolle Zusammenarbeit geht an Kristin Bunte, Luca Mao, Kurt Swingle, Heidi Schott und Elowyn Yager.


Vielen Dank an die Kraftwerke Mattmark AG (insbesondere Tibor Bilgischer u. Willi Gruber) für die Möglichkeit Forschung an der Wasserfassung des Riedbachs zu betreiben und die zur Verfügung gestellten Daten, sowie an das CESS-APUNCH Projekt (ETH), welches diese Dissertation erst ermöglicht hat.

Vielen Dank für jegliche, nicht fachliche dafür umso wertvollere, Unterstützung geht an meine Familie, an Hannah, meine Kinder Leon, Paul und Juli sowie meine Eltern und Schwiegereltern.