BEDLOAD TRANSPORT AND FLOW RESISTANCE IN STEEP CHANNELS – INTRODUCTION TO THE ISSUES IN THE CONTEXT OF MOUNTAIN BASINS OF THE CENTRAL EUROPEAN REGION

TOMÁŠ GALIA, JAN HRADECKÝ

University of Ostrava, Faculty of Science, Department of Physical Geography and Geoecology

ABSTRACT

Although sediment transport in the uppermost parts of mountain basins may be supply- limited, bedload transport in small torrents represents a potential hazard during extraordinary floods. The authors attempt to review recent worldwide research of bedload and hydraulics the results of which could potentially be applied in local steep channels. Approaches to the calculation of flow resistance (Manning relationship, logarithmic and power-law relationships) and critical conditions of the incipient motion of a bed particle (critical shear stress, stream power, flow velocity and discharge) are closely discussed in order to describe hydraulic and bedload transport conditions in steep headwater streams. Difficulties are pointed out in connection with the estimation of critical conditions of bedload transport in headwater streams with poorly sorted bed material and more or less developed bedforms. In the second part of the paper the authors present two pilot studies from the Western Carpathians and test potential applicability of specific formulas in this region. Results of the formulas point to relatively non-selective character of bedload transport during high magnitude floods in Beskydian headwater streams. An important role is played by the intensity of sediment delivery to active channels in limited sediment-supply conditions. The authors also point out that a majority of bedload transport and flow resistance equations were originally created on the basis of flume experiments or they were tested in Alpine torrents and therefore it is supposed that they might work well in local torrents after being calibrated to local conditions.

Key words: headwater stream, flow resistance, bedload transport, Western Carpathians

1. Introduction

Hydraulic and transport processes in steep channels are quite different from those occurring in low and mid-slope streams. Our study introduces locally neglected issues of bedload transport and flow resistance in steep streams and it summarizes the progress in world-wide in relation to this problematics. In addition, we present two recent research works from the Western Carpathians dealing with the transport of bed particles during 5/2010 flood in steep headwater streams. Headwater channels comprise 60–80 percent of the total length of the drainage network (Schumm, 1956). Therefore, high-magnitude floods that have occurred in the eastern part of the Czech Republic in recent years (1997, 2009, 2010) urge us to make a better assessment of their source zones. Similarly, bedload transport connected with erosion and further deposition of coarse material represents another potential hazard in mountain and piedmountain landscapes. Steep headwater channels are characterized by a wide range of sediment sizes and occurrence of large woody debris and bedrock outcrops. This fact predetermines high variation in bankfull channel geometry, longitudinal profile, flow velocity and bed roughness. Moreover, what also contributes to the formation of mountain channels is episodic colluvial processes such as debris flows and slope failures (Montgomery, Buffington 1997; Benda et al. 2005). Transport capacity during extreme flood events is very high, resulting in rapid morphological changes (Chiari, Rickenmann 2010). Sediment transport in the uppermost parts of mountain basins may be supply limited and there is often a strong interaction between hillslope processes and the channel network (Bathurst 1987; Montgomery, Buffington 1997; Chiari, Rickenmann 2010). Montgomery and Dietrich (1988) suppose that the source area above the channel head decreases with increasing local valley gradient in mountain landscapes.

A number of authors have introduced different terms of stream channels located in mountainous terrain. Wohl and Merritt (2008) consider *mountain streams* as steep (≥0.002 m/m) and confined channel segments with cobble, boulder and gravel bed substrates. Bathurst (1987), Thompson et al. (2006) describe mountain streams with a bed gradient ≥0.01 m/m. Critical bed gradient ≥0.01 m/m has been used by Chiari and Rickenmann (2010) to define *steep channels*, respectively transport conditions in them. Rickenmann and Koschni (2010) have use a term *torrent* for the most upstream reaches with a bed gradient exceeding 0.05 m/m and area of catchment ≤25 km^2, whereas Zuna (2008) has applied a bed gradient ≥0.03 m/m for Czech torrents. Originally, a term *headwater stream* was defined by Strahler (1955) as a channel of the first or second order of his classification. However, Benda et al. (2005) have pointed out difficulties in distinguishing first-order and second-order streams clearly. In their opinion, the uncertainty as for the exact definition of the first-order streams results from the concept of smooth transition

To be quoted as:
Galia, T. – Hradecký, J. (2012): Bedload transport and flow resistance in steep channels – introduction to the issues in the context of mountain basins of the Central European region
AUC Geographica, 47, No. 1, pp. 23–33

of slopes into colluvial channels. The resolution of topographic maps, particularly the evidence of gullies, should also be reflected. Halwas and Church (2002) have only recognized first-order channels of Strahler's classification as headwater channels. In the last decades, many authors have focused on developing morphological (Frissel et al. 1986; Grant et al. 1990; Hawkins et al. 1993; Halwas, Church 2002; Thompson et al. 2006) or process-based (Montgomery, Buffington 1997) classifications of either mountain channel reaches or units.

Our paper presents a comprehensive review of the issues of flow resistance and bedload transport in steep headwater streams as well as the first results of the research in the Outer Western Carpathians. Due to the complexity of the main topic, we have decided to reduce its extent and therefore the part dealing with total volumes of transported bedload material in channels (i.e. kg min^{-1}) will be presented in one of further articles. In this paper we discuss flow resistance describing the influence of bed elements on flow velocity. Critical conditions of the incipient motion of a particle of a certain diameter are introduced in the second part of the paper. Despite gradient criteria of steep channels (≥0.01 m/m), we only discuss facts connected with confined headwater channel segments. Some typically gravel-bed rivers may exceed 0.01 m/m slope, but there is no significant presence of typical bed forms influencing flow resistance and they lack channel-slope coupling. We sincerely hope that this study will bring a new insight into the observation of small mountain basins in the Czech Republic.

2. Flow resistance

The term *flow resistance* can be conceived as an influence of bed roughness on flow velocity. The higher bed roughness is, the smaller potential energy is available for sediment transport since mean flow velocity slows down. Flow resistance can be described by a well-known Manning relationship:

$$v = 1/nR^{0.67}S^{0.5}, \tag{1}$$

where v is mean flow velocity, n is Manning roughness parameter, R is hydraulic radius (m) and S is channel gradient (m/m). Owing to non-uniform character of flow in natural channels, it is more appropriate to describe parameter S as the average water-surface slope of a stream reach or the average channel gradient of a stream reach. Generally speaking, there is a tendency for n to decrease with increasing water stage until it reaches the top of the channel banks (Chiari 2008; Ferguson 2010). As headwater channels usually lack inundation area, it is very difficult to describe the tendency of n after exceeding the 'bankfull stage' without particular at-a-site measurement. Several formulas to identify n roughness parameter have been developed for mountain streams. One of the most used formulas is that of Jarett (1984), originally designed for channels of slopes 0.002 < S < 0.04 m/m, although it overestimates channel roughness (Marcus et al. 1992). It takes the form:

$$n = 0.32R^{-0.16}S^{0.38}. \tag{2}$$

In case of streams steeper than 0.008 m/m, Rickenmann (1996) introduces the following equation:

$$1/n = (0.97g^{0.41}Q^{0.19})/(S^{0.19}D_{90}{}^{0.64}), \tag{3}$$

where Q is flow discharge (m^3 s^{-1}).

Flow resistance in steep streams is traditionally divided into grain resistance and form resistance. Therefore, it cannot be described sufficiently by means of a characteristic percentile of grain size distribution (Aberle, Smart 2003). Based on the Manning roughness (n) parameter corresponding to total flow resistance or total roughness (n_{tot}), the following relationship has been derived by Rickenmann (2005) cf. Chiari and Rickenmann (2010):

$$n_r/n_{tot} = 0.092S^{-0.35}(h/D_{90})^{0.33}, \tag{4}$$

where n_r is Manning roughness coefficient associated with grain friction only and h is mean flow depth (m). Wong and Parker (2006) have suggested describing n_r as:

$$1/n_r = 23.2/\sqrt[6]{D_{90.}} \tag{5}$$

Using grain roughness after Wong and Parker (2006), Rickenmann et al. (2006) have presented an equation

$$n_r/n_{tot} = 0.133Q^{0.19}/(g^{0.96}S^{0.19}D_{90}{}^{0.47}). \tag{6}$$

The relationship containing channel slope only has been developed by Palt (2001) from field data of Himalayan mountain streams with bed gradients 0.002 < S < 0.12 m/m and it takes the form:

$$n_r/n_{tot} = 0.15S^{-0.36}. \tag{7}$$

Similarly, Chiari and Rickenmann (2010) have mentioned the following relationships based on Alpine torrents:

$$n_r/n_{tot} = 0.07S^{-0.43} \quad \text{(based on Eq. 4)} \tag{8}$$

and

$$n_r/n_{tot} = 0.07S^{-0.476} \quad \text{(based on Eq. 6).} \tag{9}$$

A large variation in the n_r/n_{tot} values is observed in channel gradients up to about 0.1. In steeper channels form roughness appears to be more important than grain roughness, as based on a limited amount of available data (Chiari 2008). Nevertheless, the latest flume experiments of Zimmermann (2010) point out that the concept of separating total resistance into grain resistance and form resistance makes no sense in shallow steep channels. He assumes that the grains themselves generate form and spill resistance.

Flow velocity approaches based on logarithmic law are probably more appropriate for shallow channels with heterogenous bed mixture (Ferguson 2010). A basic formula has been developed by Colebrook and White (1937) and it takes the form:

$$v = \sqrt{(8/f)}\sqrt{(gRS)}, \quad (10)$$

where f is the Darcy-Weisbach friction factor. In wide channels, hydraulic radius R can be substituted with mean flow depth. To describe the friction factor, Keulegan (1938) cf. Lee and Ferguson (2002) originally proposed the relationship:

$$\sqrt{(1/f)} = 2.03\log(12R/k_s), \quad (11)$$

where k_s is roughness height. In poorly sorted beds, k_s represents a D_{84} or D_{90} with a multiple in the range of 2–4 (Lee, Ferguson 2002).

Thompson and Campbell (1979) cf. Ferguson (2007) have proposed a modified Keulegan equation:

$$\sqrt{(8/f)} = 2.5(1 - 0.1k/R)\ln(12R/k), \quad (12)$$

where $k = 4.5D_{50}$ or $k = 2.37D_{84}$. Later flume experiments of Aberle and Smart (2003) led to an equation:

$$\sqrt{(8/f)} = 3.54\ln(h/D_{84}) + 4.41. \quad (13)$$

The relationship (13) is valid for low relative submergence ($h/D_{84} < 3$) and bed gradients from 0.02 to 0.1 m/m. Introducing bed shear stress τ_b (N m^{-2}), Wiberg and Smith (1991) cf. Zimmermann and Church (2001) have derived relationship for high stream gradient and high roughness:

$$v/\sqrt{(\tau_b \rho)} = (1/k)\ln(14.4h/D_{84c}), \quad (14)$$

where ρ is water density (kg m^{-3}), k is von Karman constant ~0.4 and D_{84c} indicates the shortest axis of D_{84} bed surface percentile. In case of submergence of all bed sediment, h/D_{84c} ratio is equal to 1. The relationship (14) has been used by Zimmermann and Church (2001) to express potential bed shear stress in step-pool sequences.

Finally, power law equations expressing flow velocity have been formulated by Rickenmann et al. (2006):

$$v = (1.93g^{0.5}h^{1.5}S^{0.5})/D_{90} \quad (15)$$

and by Lee and Fergusson (2002), namely on the basis of Jarrett's (1984) roughness relationship (Eq. 2):

$$\sqrt{(1/f)} = 0.35R^{0.33}S^{-0.38}. \quad (16)$$

Lee and Ferguson (2002) have tested logarithmic-law equations (Eq. 11 and Eq. 12) and a power-law equation (Eq. 16) both in flume and nature channels. The equation that worked best in all studied reaches was the power-law equation (Eq. 16). However, the Eq. 12 brought also very good results without the necessity to be calibrated. On the contrary, the Eq. 11 needed to be calibrated to the multiple $2.78D_{84}$. In addition, we suppose that D_{84} does not fully represent resistance length scale in channels with poorly sorted material and therefore we recommend optimizing the relationship for a particular channel reach. On the other hand, Bathurst (2002) sees relative submergence based on D_{84} as an excellent primary predictor of the resistance function.

Zimmermann (2010) obtained good results during his flume experiments introducing dimensionless unit discharge q^* and dimensionless velocity v^*:

$$q^* = q/(gD_{84}{}^3)^{1/2} \quad (17)$$

and

$$v^* = v/(gD_{84})^{1/2}. \quad (18)$$

Subsequently, Zimmermann (2010) expressed a relationship between q^* and v^*, which takes the form:

$$v^* = 0.79q^{*0.38}. \quad (19)$$

Based on data from an Alpine torrent Rio Cordon, Comiti et al. (2007) have derived a similar relationship:

$$v^* = 0.92q^{*0.66}. \quad (20)$$

Aberle and Smart (2003) have introduced standard deviation of bed elements elevation (σ) instead of D_{84} parameter. Although standard deviation does not fully describe the character of stepped bed (the same value can represent a few larger steps or a lot of smaller steps resulting in distinct hydraulics), Zimmermann (2010) has tested σ parameter instead of D_{84} in dimensionless velocity and discharge definitions (Eq. 17 and Eq. 18). He found out that D_{84} and standard deviation of bed elements are only poorly correlated. Nevertheless, he admitted that using σ may improve the characterization of channel roughness. The introduction of standard deviation of bed elements led to the following equation:

$$v^* = 0.63q^{*0.35}. \quad (21)$$

3. Critical conditions of the incipient motion of a bed particle

Frequent points of discussion involve selective (smaller grains are moved during lower flows) or non-selective character of bedload transport at steep slopes. Near-equimobility conditions have been found by Parker and Klingeman (1982) in gravel-bed channels and Lamarre and Roy (2008) in step-pool streams, whereas a large degree of size-selectivity has been presented by Zimmerman and Church (2001) or Lenzi et al. (2006) in step-pool channels. Intermediate conditions between the two extremes are very probable as it has been suggested by Hassan and Church (2002) and Lenzi et al. (2006). Zimmermann et al. (2010) have observed that bed stability in step-pool systems increased with decreasing channel width/D_{84} (where D_{84} is a diameter of stones building steps) for ratios less than six. Armouring of bed material and heterogeneity of bed sediments also play an important role in the definition of critical conditions that set bed grains into motion (Lenzi et al. 2006; Chiari, Rickenmann 2010).

Critical parameters, such as critical velocity, unit discharge, unit stream power and shear stress, are used to describe the incipient motion of a particle of a certain diameter (b-axis length). It is particularly the concept of critical shear stress that has often been mentioned in recent literature (Zimmermann and Church 2001, Mueller and Pitlick 2005, Lenzi et al. 2006, Mao et al. 2008) and the basic relationship takes the form:

$$\tau_{ci} = (\rho_i - \rho)\tau_{ci}{}^{*}D_i, \quad (22)$$

where τ_{ci} (N m^{-2}) is critical shear stress acting on grain D_i (m), ρ_i is grain density and $\tau_{ci}{}^{*}$ is dimensionless critical shear stress or the so-called Shields parameter of a representative grain size (D_i in our case). Since $\tau_{ci}{}^{*}$ was introduced by Shields (1936), uncertainty in the definition of $\tau_{ci}{}^{*}$ exact value has remained unresolved. Shields (1936) has recommended a value of 0.045 for heterogeneous bed mixture, whereas Zimmermann and Church (2001) have used values of 0.03, 0.045, and 0.06 for step-pool streams and D_{84} diameter. Buffington and Montgomery (1997) have pointed out that there is no definitive $\tau_{ci}{}^{*}$ value for D_{50} for gravel-bed rivers and they refer to an ascertained range of 0.03–0.086. Mueller and Pitlick (2005) have set up $\tau_{ci}{}^{*}$ values to 0.06–0.12 for bed gradients of the range of 0.02–0.05 m/m. The problematics of dimensionless critical shear stress in steep channels has closely been discussed by Lenzi et al. (2006). On the basis of the research in Rio Cordon step-pool stream he has introduced relationships between D_{50} or D_{90} and D_i:

$$\tau_{ci}{}^{*} = 0.143(D_i/D_{50})^{-0.737} \quad (23)$$

and

$$\tau_{ci}{}^{*} = 0.054(D_i/D_{90})^{-0.737}. \quad (24)$$

Prior to that, the same-based relationship was found by Andrews (1983) for the diameter of D_{50}:

$$\tau_{ci}{}^{*} = 0.0834(D_i/D_{50})^{-0.832}. \quad (25)$$

Based on the research of Australian upland streams, Thompson and Croke (2008) have derived the relationship:

$$\tau_{ci}{}^{*} = 0.068(D_i/D_{50})^{-0.6}. \quad (26)$$

It is important to note that the Eq. 24 calculating with D_{90} parameter seems to be more appropriate for boulder-bed streams because of the importance of grain structures built up by larger stones (Lenzi et al. 2006).

On the contrary, Lamb et al. (2008) have presumed a strong dependence of slopes on dimensionless shear stress. His flume experiments led to the relationship:

$$\tau_{ci}{}^{*} = 0.15S^{0.25}. \quad (27)$$

Later, Parker et al. (2011) established a similar power relation:

$$\tau_{ci}{}^{*} = 0.19S^{0.28}. \quad (28)$$

Another critical parameter, unit stream power, is expressed in the form:

$$\omega = (\rho QSg)/w, \quad (29)$$

where w is channel width (m) which is appropriate in case of known discharge. Some authors (Williams 1983; Petit et al. 2005; Mao et al. 2008) have published equations calculating a critical value of unit stream power which is able to move a particle of a certain diameter. Williams (1983) derived his relationship after observing worldwide gravel-bed streams. Lower (Eq. 30) and upper (Eq. 31) limits of such a critical condition take the forms:

$$\omega_{ci} = 0.079D_i^{1.3} \text{ for } 10 < D_i < 1500, \quad (30)$$

$$\omega_{ci} = 2.9D_i^{1.3} \text{ for } 15 < D_i < 500 \; (D_i \text{ in mm}). \quad (31)$$

For Belgian gravel-bed rivers, i.e. not headwater channels, Petit et al. (2005) have presented the relationship:

$$\omega_{ci} = 0.13D_i^{1.438} \text{ for } 20 < D_i < 150 \; (D_i \text{ in mm}). \quad (32)$$

For grains larger than 50 mm in diameter, Costa (1983) has defined an equation based on data from gravel-bed streams of Colorado, USA:

$$\omega_{ci} = 0.03D_i^{1.686}. \quad (33)$$

Investigating Alpine and Andean environments and step-pool channels Mao et al. (2008) have established a relationship for a wide range of particle diameters (c. 2–1000 mm):

$$\omega_{ci} = 31.502D_i^{0.488}. \quad (34)$$

On the ground of small flood accumulations in local Beskydian headwater streams, a relationship has been derived for $200 < D_i < 400$ mm only (Galia, Hradecký 2011).

$$\omega_{ci} = 3.0397D_i^{0.7885}. \quad (35)$$

As mentioned above, there are a number of slightly different relationships for unit stream power. Local channel parameters such as bed and form roughness along with the imbrication of bed sediments play an important role in the assessment of the value of critical unit stream power for a grain of a certain diameter.

Although critical velocity can physically be defined easily, there are difficulties in measuring this parameter in steep channels with prevailing turbulent and near critical flows. Flow velocities are usually significantly different between steps and pools (e.g. Chin 2003; Wilcox, Wohl 2007). Nevertheless, some works dealing with critical stream velocity have been published. One of frequently cited studies is that by Costa (1983), who reconstructed palaeofloods and critical conditions that, in some cases, set into motion extraordinarily large stones (up to 3.2 m) in the Colorado region. The relationship takes the form:

$$v_{ci} = 0.18D_i^{0.487}, \quad (36)$$

where v_{ci} is a critical velocity acting on a diameter D_i (mm). Costa (1983) recommended using the Eq. 36 to compute an average flow velocity in case intermediate axes of the five largest boulders moved by the flood are known. (e.g. five largest boulders creating a step).

For worldwide gravel-bed streams, Williams (1983) has derived similar relationships with the upper (Eq. 37) and lower (Eq. 38) boundaries:

$$v_{ci} = 0.46D_i^{0.5} \quad (37)$$

and

$$v_{ci} = 0.065D_i^{0.5}. \quad (38)$$

Eq. 37 is valid for a range $15 < D_i < 500$ mm and Eq. 38 for grain diameters $10 < D_i < 1500$ mm.

Finally, the parameter of critical unit discharge q_{ci} acting on a grain of a diameter D_i (m) has been used by several authors since Schoklitsch (1962), who derived the relationship during flume experiments:

$$q_{ci} = 0.15g^{0.5}D_i^{1.5}S^{-1.12}, \quad (39)$$

which is valid for slopes $0.0025 < S < 0.2$ m/m and the b-axis length of a grain $3 < D_i < 44$ mm. Bathurst (1987) has used a simple relationship:

$$q_{ci} = aD_i^{b} \quad (40)$$

and fitted coefficients a (0.0933–0.168) and b (0.201–0.392) to each reference reach and a certain particle diameter (68–94 mm). Furthermore, to add hiding/exposure effect, Bathurst (1987) has expressed a relationship for b coefficient:

$$b = 1.5(D_{84}/D_{16})^{-1}. \quad (41)$$

Similarly, Lenzi et al. (2006) have developed a relationship for an Alpine stream Rio Cordon:

$$q_{ci} = 1.176D_i^{0.641}. \quad (42)$$

However, their b coefficient does not fit Bathurst's (1987) defined interval. The relationship (Eq. 42) is valid for a wide range of grain diameters $0.04 < D_i < 1$ m.

Using dimensionless critical discharge q^*_{ci} similar to Eq. (17) and introducing D_{90} instead D_{84}, Lenzi et al. (2006) mentioned a significant correlation between q^*_{ci} and a particle diameter D_i:

$$q^*_{ci} = 0.745(D_i/D_{90})^{-0.859}. \quad (43)$$

In contrast to critical parameters necessary to move a grain of a certain diameter, there are only few published papers dealing with the transport distance of an individual grain in small channels. The application of magnets, iron bolds or transmitters implanted into coarse grains has facilitated the determination of a change in the position of a grain during various flow discharges (Ergenzinger, Schmidt 1990; Lenzi 2004). Ergenzinger and Schmidt (1990) have studied the relationship of particle weight (0.33–5.02 kg) and travel distance in step-pool Alpine torrents and discovered a weak tendency to shorter travel distance of the heaviest particles only. Transport distance of lighter grains did not correlate with their weights. It was also found out that disk-shaped pebbles had three-four times shorter travel distance than rods, balls and ellipsoids. On the other hand, revising bedload traps in a British headwater stream Demir and Walsh (2005) revealed that discs are preferentially entrained at lower flows, closer to the threshold for motion.

For a gravel-bed pool-riffle stream Hassan et al. (1992) have developed an equation that led to the identification of travel distance length L_i (m) of a grain diameter D_i, if D_{50} travel distance length L_{50} is known:

$$L_i/L_{50} = 1.77[(1 - \log_{10}(D_i/D_{50sub})]^{1.35}, \quad (44)$$

where D_{50sub} is a diameter of a median grain size of subsurface bed material. Eq. (44) is therefore useful if there are no available data on peak flood discharges. Moreover, Lenzi et al. (2004) have proposed the following equations for an Alpine step-pool stream:

$$L_i = 17.72(Q_p - Q_{ci})^{1.01} \quad (45)$$

and

$$L_i = 4.035(Q_p - Q_{ci})^{1.7}, \quad (46)$$

where Q_p is peak discharge and Q_{ci} critical discharge necessary to move a grain of a certain diameter (m). Parameter Q_{ci} can be obtained easily by means of some of the equations used to compute unit critical discharge for a particle (Eq. 39–43).

4. Case studies from the West-Carpathian region

4.1 Critical parameters setting in motion the largest boulders in a Beskydian headwater stream during the 5/2010 flood

In May 2010 an extraordinary flood affected the region of the north-eastern part of the Czech Republic. Galia and Hradecký (2011) focused on accumulations that developed in small Beskydian headwater channels during this event. The aim of that study was to determine critical parameters of flow causing the transport of the largest boulders, which consequently settled down forming accumulations. Nine locations with such flood deposits were discovered in Lubina and Malá Ráztoka basins. All the three axis of five largest boulders from the surface of each accumulation were measured. The authors presumed that the boulders had passed channel-reaches above store areas; therefore geometrical parameters of flood channels were taken above the accumulations and subsequently particle-size analysis of surface-bed material was obtained. In conclusion, the gradient of channel-reaches varied between 0.06–0.14 m/m, the flood channel width between 2.5–7 m, flow depth between 0.3–0.45 m and D_{90} between 0.120–0.165 m. According to Montgomery and Buffington's classification (1997), bedrock-cascade, cascade, step-pool and plane bed channels were observed. Peak flood discharge data were provided by Malá Ráztoka gauging station (VÚLHM, Frýdek-Místek branch). The peak discharge reached almost 4 $m^3\ s^{-1}$ at Malá Ráztoka gauging station (2.01 km^2), which corresponded to 25-year flood discharge derived from 1954 to 2003 time series. For studied channel-reaches and basin areas, peak discharges were computed in the range of 0.4–2.8 $m^3\ s^{-1}$ using a specific simple discharge method.

The authors made an effort to compute critical velocity, shear stress, unit stream power and unit discharge, which had an effect on the movement of the largest boulders and their later deposition in channel accumulations during a c. Q_{25} flood. Firstly, in order to obtain values of local dimensionless shear stress, Eq. (24) presented by Lenzi et al. (2006) was used. To make a comparison, a slope-based relationship (Eq. 27) of Lamb et al. (2008) was also applied. In the next step, values of critical shear stress were simply obtained by means of the relationship of Eq. 22. Applying the approach of Lenzi et al. (2006) the obtained result showed that there was a relatively small difference in critical values of the largest boulders ($0.2 < D_i < 0.4$ m), which varied between 115–160 N m^{-2}. Dimensionless shear stress, obtained using the same method, reached the values of 0.026–0.038 for the largest boulders, which did not correspond to the original Shields values (0.045 for heterogeneous bed sediment). In contrast, values of the same parameter obtained using the relationship of Lamb et al. (2008) led to total overestimation (0.078–0.092) and corresponded to D_{50}–D_{90} values obtained using the method of Lenzi et al. (2006). Therefore, in case of relative homogenous bed material the equation of Lamb et al. (2008) is generally appropriate in steep slopes, but in case of individual large boulders we highly recommend the approach of Lenzi et al. (2006) based on D_i/D_{90} relationship.

As it was mentioned above, the estimation of critical velocity is problematic in steep gradients. Rickenmann's (2006) power-law equation (Eq. 15) brought surprisingly unrealistically high values of flow velocity (up to 4 m s^{-1} in narrow channels). Introducing critical shear stress values obtained by Lenzi et al. (2006) approach into Eq. 14 led to quite uniform flow velocity values in a range of 1.36 to 1.68 m s^{-1} for boulder diameters $0.2 < D_i < 0.4$ m. This could point to less size-selective transport in headwater channels.

The authors also examined the relationship between unit critical discharges, unit stream power and maximal diameters of transported boulders. No dependent relationships were found between these parameters. A newly created Eq. 35 showed only insignificant correlation ($R^2 = 0.053$) between unit stream power and a maximal diameter of stored boulders $0.2 < D_i < 0.4$ m. This again denotes less size-selective bedload transport during a high-magnitude flood. As shown in Fig. 1, a similar trend occurs between relationships presented by locally derived Eq. 35, Eq. 34 of Mao et al. (2008) and Eq. 30 of Williams (1983).

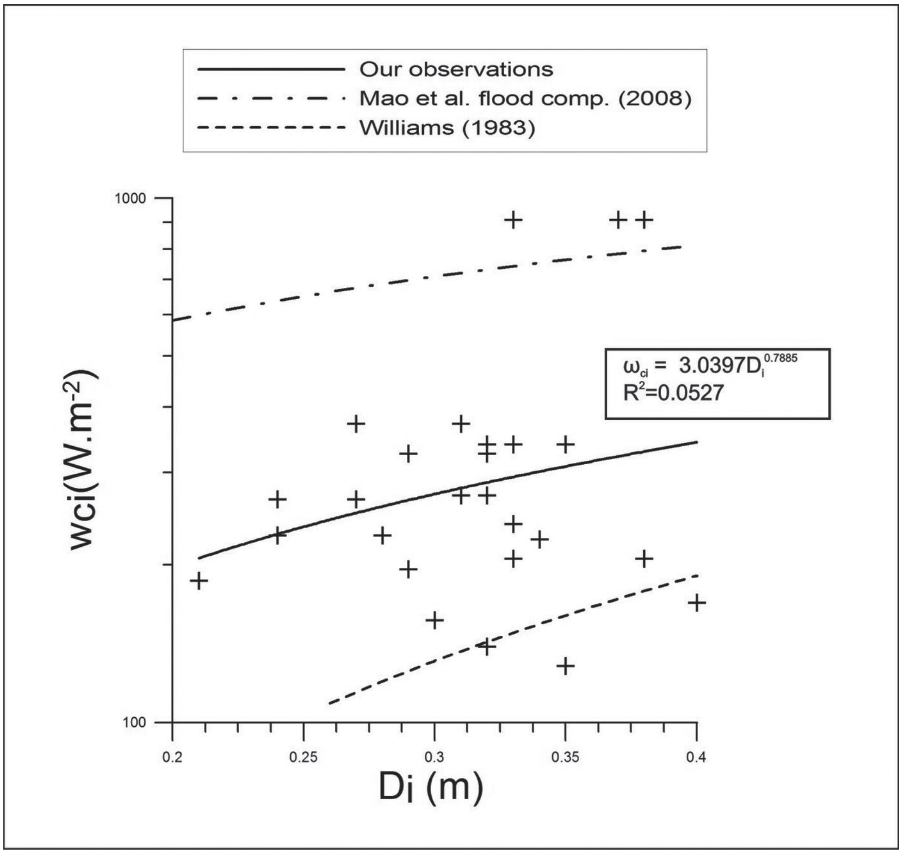

Fig. 1 Relationship between D_i of stored boulders and unit stream power. Observed data were compared with trends of Eq. 30 and Eq. 34

Moreover, any correlation (R^2 = 0.025) was identified between unit discharge and a boulder diameter. The intensity of bedload transport in high gradient streams during floods is rather dependent on actual sediment delivery, namely because of sediment-supply limited character of headwater basins (Montgomery, Buffington 1997, Yu et al. 2010). For more details on the study see Galia, Hradecký (in press).

4.2 Transport distance of bed material in mid-mountain headwater streams

During the winter season of 2009–2010 a small check-dam was built using limestone boulders and the limestone fraction was used to fortify an adjacent forest road in Libotínský potok basin (total A ~ 3 km²). This small stream located in Beskydian piedmont (300–450 m a.s.l.) is geologically composed of Těšín-Hradiště Formation with the occurrence of claystone and sandstone layers interlaid with teschenite igneous rocks. This fact points to the possibility to monitor limestone fraction in the channel and measure the transport distance of individual grains from a certain point, i.e. check-dam with an adjacent forest road located at river km 1.7 (A = 1.12 km²) in our case. We presume that the transport was most intensive during the 5/2010 flood, but naturally some previous smaller flood events occurring between the winter season of 2009/2010 and May 2010 had effect on travel distance of particles as well. Unfortunately, no discharge data are available from the basin; therefore, we have tried

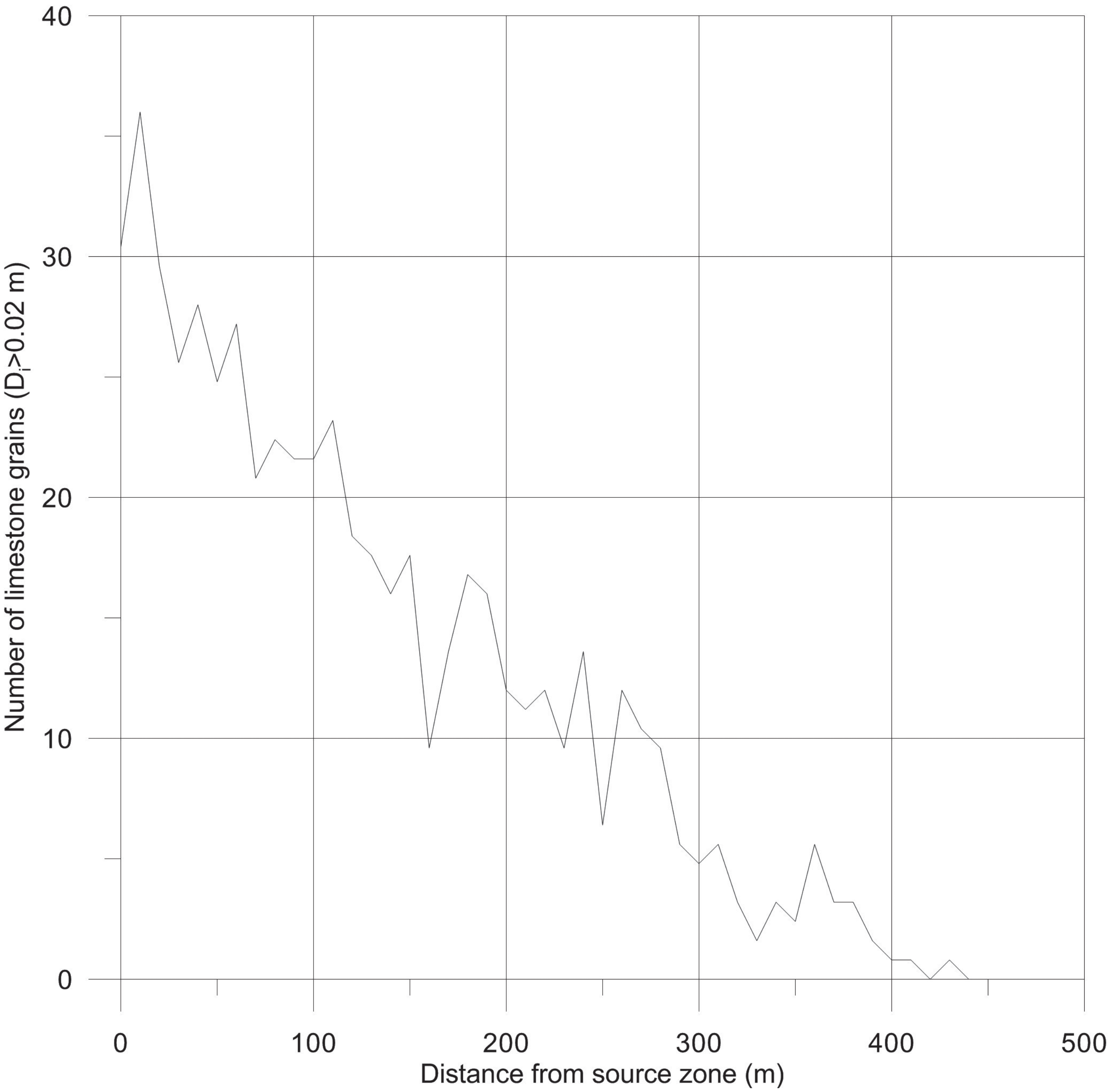

Fig. 2 Number of limestone particles per m²

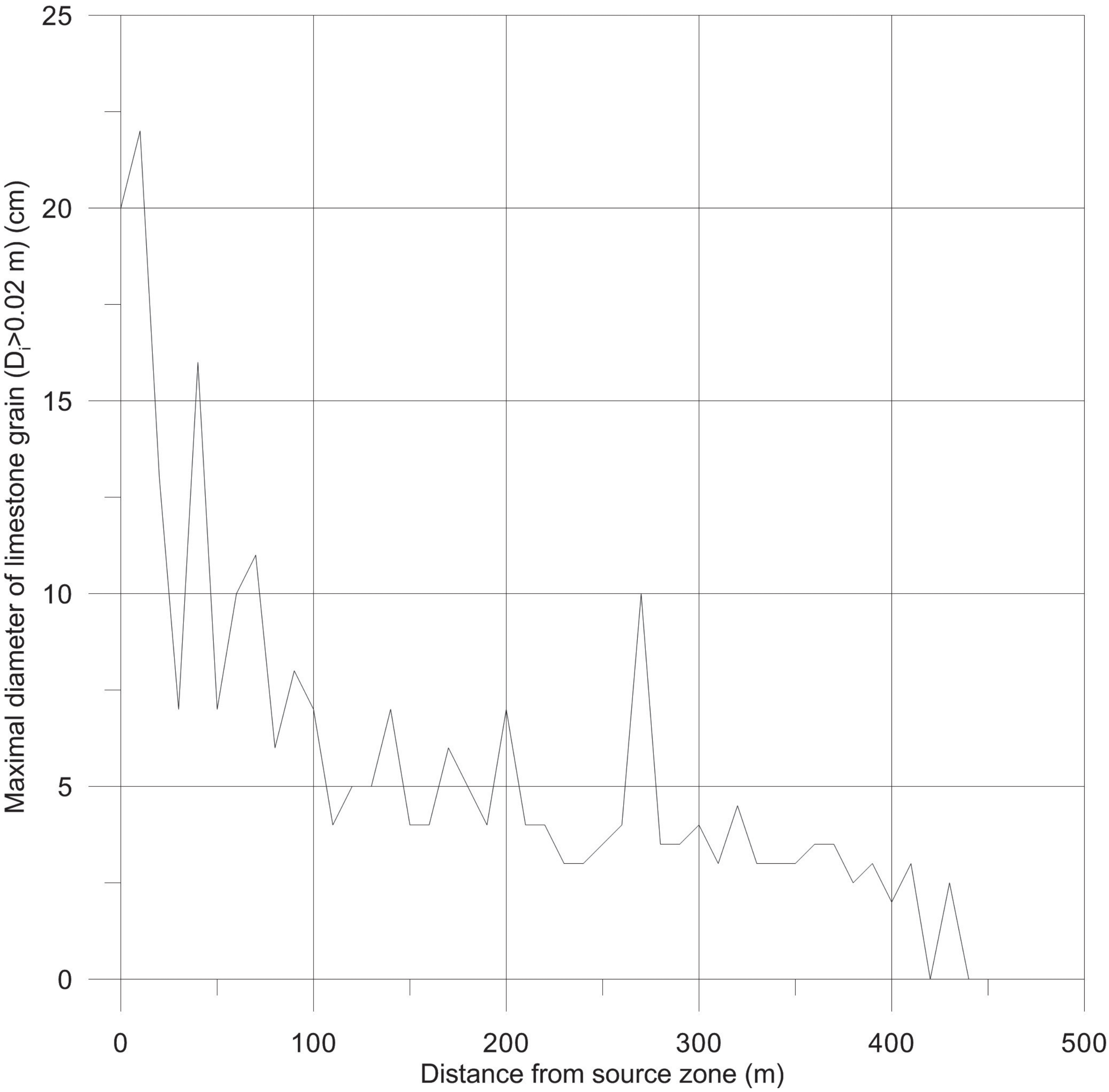

Fig. 3 Maximal observed diameter of a limestone grain

to reconstruct the peak 5/2010 flood discharge on the basis of channel geometric characteristics and Manning roughness parameter.

We have counted the amount of limestone particles (min $D_i > 0.02$ m) at the distance of every 10 m of the stream longitudinal profile (adjacent area ±0.5 m) and have measured widths of the bankfull channel (2.8–3.5 m). This way we obtained the amount of limestone particles per unit channel width. We also measured *a*, *b*, and *c* axes of the largest grain in each reach. The channel is characterized by rapid flow (plane bed) character at 1.7–1.9 river km (0.04–0.07 m/m channel gradient), then further down it changes gradually into a pool-riffle channel (about 0.03–0.05 m/m). D_{50} varies between 0.045 m for rapid channel reaches and 0.035 m for pool-riffle channel reaches, respectively. Similarly, values of 0.11 m and 0.065 m were obtained for D_{90}. Manning roughness parameter was computed for a bankfull channel using Jarrett's (1984) method (Eq. 2), whereas in case of a rapid channel its value was 0.141 and in case of a pool-riffle channel it was 0.122. The values, which were obtained on the basis of average parameters of observed rapid and pool-riffle channel morphology, are only approximate. We included the roughness parameter in Rickenmann's (1996) relationship for discharge (Eq. 3) and computed theoretical discharge during the bankfull stage (Q_1–$Q_{2.5}$). However, the resulting value of about 0.002 $m^3\ s^{-1}$ in a pool-riffle reach seems to be underestimating and it shows us somewhat problematic use of Eq. 2 or rather Eq. 3 for steep streams with a relatively low D_{90} and thus uniform bed sediments. Reading high-level water marks inside the channel and adjacent floodplain it was found out that the 5/2010 peak discharge exceeded the bankfull stage. We estimated that it could be about 20y discharge, most likely in values of the range of 1–2 $m^3 s^{-1}$, namely because peak discharge in Jičínka stream reached Q_{20} stage at Nový Jičín gauging station (76 km^2; 70 $m^3 s^{-1}$; 6 km far from our study area) during the 5/2010 flood event.

In the source zone of limestone particles, we counted 0–50 particles $D_i > 0.02$ m per m^2. 70 meters far from the main source, the amount of grains did not exceed 30 particles per m^2. The number of particles decreased to 10–20 particles per m^2 between the 200th–280th m and as far as the 430th m from the source zone the amount of particles did not reach 10 pebbles per m^2. The trend in the amount of particles *A* per m^2 per travel distance *L* (m) is linear and it takes the form:

$$A = -0.745L + 29.711 \; (R^2 = 0.93) \qquad (47)$$

Since no limestone particles were observed further down the stream, the maximal competence of Q_{20} flood in the described headwater stream is to transport a grain $D_i > 0.02$ m as far as the 440th m of the distance (Fig. 2). Relatively more particles were deposited in small benches in rapid channels than in straight channels void of accumulation forms or pools.

B axis of the largest observed cobble, which was located in the source zone of sediments, was 0.22 m long. Within the studied reach no limestone grain of $D_i > 0.1$ m was observed at the distance of 80–440 m far from the source zone with one exception at the 270th m (Fig. 3). The best-fitting trend of limestone particle diameter D_i ($0.02 < D_i < 0.22$ m) and travel distance *L* (m) takes the form:

$$D_i = 27.44L^{-0.568} \; (R^2 = 0.77) \qquad (48)$$

The 5/2010 flood destroyed the check dam at 1.7 river km where slightly moved boulders of approximately 0.5 m in diameter were observed (Fig. 4). Involving the 0.5 m diameter in the Costa's (1983) relationship for flow velocity (Eq. 36), we obtained values of c. 3.7 m s^{-1}, which seem to be unrealistic.

In contrast, using the equation of Lenzi et al. (2006) to obtain critical unit discharge (Eq. 42), we got 0.75 $m^2 s^{-1}$ value of q_{ci} for 0.5 m boulder diameter. The rapid channel is about 3 m wide in the studied reach; therefore the resulting peak discharge value of 2.25 $m^3 s^{-1}$ seems to be only a little overestimated. We suppose there could be great differences in critical parameters of grains in well armoured channel-beds and grains from sediment sources or, in our case, material imported into the channel in connection with human activity.

5. Conclusion

This paper summarizes diverse attempts to compute mean flow velocity in mountain headwater channels based on flow resistance approaches (Manning relationship, logarithmic and power-law relationships). Resulting flow velocity values for local streams may differ due to the character of flume experiments and different environments for which most of the equations were derived. In spite of this fact we suppose that most of the flow resistance based equations may work well after being calibrated to local conditions. Some authors (Ferguson 2010) recommend not using the basic Manning equation due to variations in roughness parameter with increasing water stage; however, this relationship is still widely used in hydrological modeling and it has even been applied in recent routing sediment models for steep channels (Rickenmann et al. 2006; Chiari, Rickenmann 2010).

Furthermore, we present several methods to obtain values of critical parameters necessary to move a bed particle during high flows. Despite the discrepancy between theories of selective and non-selective bedload transport, the terms of critical shear stress and unit stream power are often used to describe such critical conditions for individual grains. It is necessary to estimate values

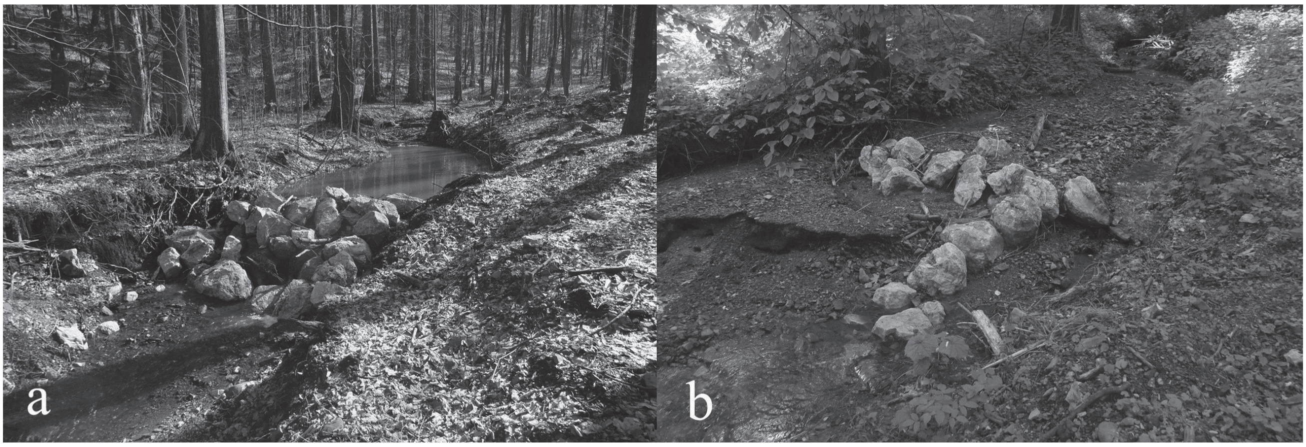

Fig. 4 Check dam at 1.7 river km of Libotínský potok – a) before the 5/2010 flood, b) after the 5/2010 flood

of dimensionless shear stress as exactly as possible due to heterogeneity of bed mixture and possible stable bed structures (e.g. jamming steps, armouring of bed) (Lenzi et al. 2006; Zimmermann et al. 2010).

The case studies point out the importance of further research in small Carpathian mountain basins that often contain relatively less coarse bed surface material than Alpine torrents, for which many flow resistance and bedload transport equations have been developed. Some relationships presented in our paper are not strictly valid for headwater streams of the character presented above and it is necessary to verify them and adjust them to local conditions. The main topic of future studies should also be better assessment of source zones of sediments as an important factor determining the character and intensity of bedload transport in supply-limited zones.

Acknowledgements

The research was supported by the Student Competition Grant No. SGS6/PřF/2011. The authors also express sincere thanks to Monika Hradecká for English language revision.

REFERENCES

ABERLE, J., SMART, G. (2003): The influence of roughness structure on flow resistance on steep slopes. Journal of Hydraulic Research 41, pp. 259–269.
BATHURST, J. C. (1987): Critical conditions for bed material movement in steep, boulder-bed streams. Erosion and Sedimentation in the Pacific Rim (Proceedings of the Corvallis Symposium, August, 1987), pp. 309–318. IAHS Publ. no. 165.
BATHURST, J. C. (2002): At-a-site variation and minimum flow resistance for mountain rivers. Journal of Hydrology 269, pp. 11–26.
BENDA, L., HASSAN, A., CHURCH, M., MAY, C. L. (2005): Geomorphology of steepland headwaters: The transition from hillslopes and channels. Journal of the American Water Resources Association 41, pp. 835–851.
BUFFINGTON, J. M., MONTGOMERY, D. R. (1997): A systematic analysis of eight decades of incipient motion studies, with special reference to gravel-bedded rivers. Water resources research 33, pp. 1993–2029.
CHIN, A. (2003): The geomorphic significance of step-pools in Mountain streams. Geomorphology 55, pp. 125–137.
COLEBROOK, C. F., WHITE, C. M. (1937): Experiments with Fluid Friction in Roughened Pipes. Proceedings of the Royal Society of London. Series A, Mathematical and Physical Sciences 161 (906), pp. 367–381.
COMITI, F., MAO, L., WILCOX, A., WOHL, E. E., LENZI, M. A. (2007): Field derived relationship for flow velocity and resistance in high-gradient streams. Journal of Hydrology 340, pp. 48–62.
DEMIR, T., WALSH, R. P. D. (2005): Shape and size characteristics of bedload transported during winter storm events in the Cwm Treweryn stream, brecon Bacon, South Wales. Turkish Journal of Earth Sciences, Vol. 14, pp. 105–121.
ERGENZINGER, P., SCHMIDT, K.-H. (1990): Stochastic elements of bed load transport in a step-pool mountain river. Hydrology in Mountainous Regions. II – Artificial Resenvirs; Water and Slopes (Proceedings of two Lausanne Symposia, August 1990), pp. 39–46. IAHS Publ. no. 194.
FERGUSON, R. I. (2007): Flow resistance equations for gravel-and boulder-bed streams. Water Resource Research 43, W05427, doi: 10.1029/2006WR005422.
FERGUSON, R. (2010): Time to abandon the Manning equation? Earth Surface Processes and Landforms 35, pp. 1873–1876.
GALIA, T., HRADECKÝ, J. (2011): Bedload transport and morphological effects of high-magnitude floods in small headwater streams – Moravskoslezské Beskydy Mts. (Czech Republic). Journal of Hydrology and Hydromechanics 59, 4, pp. 238–250.
HALWAS, K. N., CHURCH, M. (2002): Channel units in small, high gradient streams on Vancouver Island, British Columbia. Geomorphology 43, pp. 243–256.
HASSAN, M. A., CHURCH, M., ASHWORTH, P. J. (1992): Virtual rate and mean distance of travel of individual clasts in gravel-bed channels. Earth Surface Processes and Landforms 17, pp. 617–627.
CHIARI, M. (2008): Numerical modelling of bedload transport in torrents and mountain streams. Ph.D. Thesis, Institute of mountain risk engineering, University of natural resources and applied life science, Wienna, 212 p.
CHIARI, M., RICKENMANN, D. (2011): Back-calculation of bedload transport in steep channels with a numerical model. Earth Surface Processes and Landforms 36, 6, pp. 805–815.
JARRETT, R. D. (1984): Hydraulics of high-gradient rivers. Journal of Hydraulic Engineering 110(11), pp. 1519–1539.
LAMB, M., DIETRICH, W., VENDITTI, J. (2008): Is the critical shear stress of incident sediment motion dependent on channel-bed slope? Journal of Geophysical Research-Earth Surface 113. F02008, doi: 10.1029/2007JF000831.
LAMMARE, H., ROY, A. G. (2008): The role of morphology on the displacement of particles in a step-pool river system. Geomorphology 99, pp. 270–279.
LEE, A. J., FERGUSON, R. I. (2002): Velocity and flow resistance in step-pool streams. Geomorphology 46, pp. 59–71.
LENZI, M. A. (2004): Displacement and transport of marked pebbles, cobbles and boulders during floods in a steep mountain stream. Hydrological processes 18, pp. 1899–1914.
LENZI, M. A., MAO, L., COMITI, F. (2006): When does bedload transport begin in steep boulder-bed stream? Hydrological processes 20, pp. 3517–3533.
MARCUS, W. A., ROBERTS, K., HARVEY, L., TACKMAN, G. (1992): An evaluation of methods for estimating Manning's n in small mountain streams. Mountain Research and Development 12, pp. 227–239.
MAO, L., UYTTENDAELE, G. P., IROUMÉ, A., LENZI, M. A. (2008): Field based analysis of sediment entrainment in two high gradient streams located in Alpine and Andine environments. Geomorphology 93, pp. 368–383.
MONTGOMERY, D. R., DIETRICH, W. E. (1988): Where do channels begin? Nature 336, pp. 232–234.
MONTGOMERY, D. R., BUFFINGTON, J. M. (1997): Channel reach morphology in mountain drainage basins. Geological Society of America Bulletin 109, pp. 596–611.
MUELLER, E. R., PITLICK, J. (2005): Morphologically based model of bed load transport capacity in a headwater stream. Journal of geophysical research 110, F02016, doi: 10.1029/2003JF000117.
PALT, S. (2001): Sedimenttransporte im Himalaya-Karakorum und ihre Bedeutung für Wasserkraftanlagen (in German).

Dissertation thesis, Instituts für Wasserwirtschaft und Kulturtechnik, Universität Karlsruhe.
PARKER, C., KLINGEMAN, P. C. (1982): On why gravel bed streams are paved. Water Resources Research 18(5), pp. 1409–1423.
PARKER, C., CLIFFORD, N. J., THORNE, C. R. (2011): Understanding the influence of slope on the threshold of coarse grain motion: Revisiting critical stream power. Geomorphology 126, pp. 51–65.
RICKENMANN, D. (1996): Fliessgeschwindigkeit in Wildbächen und Gebirgsflüssen (in German). Wasser, Energie, Luft 88(11/12), pp. 298–304.
RICKENMANN, D., CHIARI, M., FRIEDL, K. (2006): SETRAC – A sediment routing model for steep torrent channels. In R. Ferreira, E. A. J. Leal, and A. Cardoso (Eds.): River Flow 2006, Vol. 1, London, pp. 843–852. Taylor & Francis.
RICKENMANN, D., KOSCHNI, A. (2010): Sediment loads due to fluvial transport and debris flows during the 2005 flood events in Switzerland. Hydrological Processes 24, pp. 993–1007.
SCHUMM, S. A. (1956): Evolution of Drainage Systems and Slopes in Badlands at Perth Amboy, New Jersey. Bulletin of the Geological Society of America 67, pp. 597–646.
SHIELDS, A. (1936): Anwendung der Aehnlichkeitsmechanik der Turbulenzforschung auf die Geschiebebewegung. Mitt. Preuss. Versuchsanst. Wasserbau Schiffbau 26, pp. 5–24. English translation by W. P. Ott and J. C. van Uchelen, 43 p., United States Dept. Agriculture Soil Conservation Service Cooperative Laboratory.
SCHOKLITSCH, A. (1962): Handbuch des Wasserbaues, 3rd edn. Springer, Wien.
STRAHLER, A. N. (1955): Quantitative analysis of watersheed geomoprhology. Transaction of the American Geophysical Union 38, pp. 913–920.
THOMPSON, C. J., CROKE, J., OGDEN, R., WALLBRINK, P. (2006): A morpho-statistical classification of mountain stream reachtypes in southeastern Australia. Geomorphology 81, pp. 43–65.
THOMPSON, C. J, CROOKE, J. (2008): Channel flow competence and sediment transport in upland streams in southeast Australia. Earth Surface Processes and Landforms 33, pp. 329–352.
WILCOX, A. C., WOHL, E. E. (2007): Field measurements of three-dimensional hydraulics in a step-pool channel. Geomorphology 83, pp. 215–231.
WILLIAMS, G. P. (1983): Paleohydrological methods and some examples from Swedish fluvial environments. Geografiska Annaler 65A, pp. 227–243.
WOHL, E. E., MERRITT, D. M. (2008): Reach-scale channel geometry of mountain streams. Geomorphology 93, pp. 168–185.
WONG, M., PARKER, G. (2006): Reanalysis and Correction of Bed-Load Relation of Meyer-Peter and Müller Using Their Own Database. Journal of Hydraulic Engineering 132, 1159, doi: 10.1061/(ASCE)0733-9429(2006)132:11(1159).
YU, G., WANG, Z., ZHANG, K., CHANG, T., LIU, H. (2009): Effect of incoming sediment on the transport rate of bed load in mountain streams. International Journal of Sediment Research 24, pp. 260–273.
ZIMMERMANN, A. (2010): Flow resistance in steep streams: An experimental study. Water Resources Research 46, W09536, doi: 10.1029/2009WR007913.
ZIMMERMANN, A., CHURCH, M. (2001): Channel morphology, gradient profiles and bed stresses during flood in a step-pool channel of mountain streams. Geomorphology 40, pp. 311–327.
ZIMMERMANN, A., CHURCH, M., HASSAN, M. A. (2010): Step-pool stability: Testing the jammed state hypothesis. Journal of geophysical research 115, F02008, doi: 10.1029/2009JF001365
ZUNA, J. (2008): Hrazení bystřin (in Czech). Česká technika – nakladatelství ČVUT, Praha. 180 p.

RÉSUMÉ

Dnový transport sedimentů a drsnost koryt bystřinných toků – úvod do problematiky v kontextu horských povodí středoevropského regionu

Dnový transport sedimentů v bystřinách představuje potenciální riziko během mimořádných povodňových událostí i přes často omezenou dotaci sedimentů do horských segmentů toků. Autoři přistoupili ke zhodnocení současného stavu poznání problematiky hydrauliky bystřinných toků a kritických podmínek pro počátek transportu dnových sedimentů v těchto tocích. Podrobněji jsou popsány přístupy vyjádření vztahu drsnosti koryt vzhledem k rychlosti proudění (Manningův vztah, logaritmický a mocninný rychlostní profil) a kritické podmínky pro uvedení klastu o určitých rozměrech do pohybu (kritické tečné napětí, jednotkový výkon toku, rychlost proudění a jednotkový průtok). Vzhledem k velké variabilitě koryt vysokogradientových toků, ať už z hlediska heterogenity dnových sedimentů nebo výskytu dnových forem (stupně, tůně), je obtížné tyto kritické podmínky přesně determinovat. Ve druhé části textu jsou prezentovány dvě pilotní studie z oblasti Západních Karpat, ve kterých byly mimo jiné testovány některé z výše uvedených vztahů a zhodnoceno jejich použití pro tento region. Bylo zjištěno, že většina testovaných rovnic může být uplatněna ve zdejších bystřinách, byť některé z nich je třeba přizpůsobit místním podmínkám. Výsledky rovněž poukázaly na relativně nezávislý charakter dnového transportu sedimentů z hlediska rozměrů přeplavených valounů a balvanů během mimořádných povodní, kdy i velmi drobný tok (A < 1 km^2) dokáže během takové události transportovat téměř všechny velikostní frakce dnového materiálu. Velmi významný je i vliv přímé donášky materiálu do aktivního koryta vzhledem k faktu, že horské segmenty toků jsou většinou chápány jako systémy s nízkou dotací sedimentů.

Tomáš Galia, Jan Hradecký
University of Ostrav, Faculty of Science
Department of Physical Geography and Geoecology
Chittussiho 10
710 00 Ostrava
Czech Republic
E-mail: tom.galia@seznam.cz, jan.hradecky@osu.cz