ESTIMATION OF BEDLOAD TRANSPORT IN HEADWATER STREAMS USING A NUMERICAL MODEL (MORAVSKOSLEZKÉ BESKYDY MTS, CZECH REPUBLIC)

TOMÁŠ GALIA, JAN HRADECKÝ

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

ABSTRACT

The TOMSED numerical model, originally developed for Alpine torrents, was successfully applied on the May, 2010 flood event in headwater streams of the Czech part of the Western Flysch Carpathians. The Manning equation was used together with application of a reduced energy gradient and potential erosion limits. Contributions of form resistance and grain resistance to Manning's roughness parameter were considered separately. The absence of bedload discharge measurement made it difficult to validate the model for local headwater channels, although channel reaches with recent erosional and depositional trends were recognized fairly well by the simulations. Bedload transport was calculated in the range from 480–1240 m^3 for the richly-supplied headwater stream and 380–860 m^3 for the stream with limited sediment-supply conditions. It turns out that the channel reaches with higher sediment supply have less developed bedforms (e.g. steps) and thus lower flow resistance resulting in a higher intensity of bedload transport.

Keywords: headwater stream, hydrological modelling, bedload transport, TOMSED, Moravskoslezské Beskydy Mts

1. Introduction

Steep headwater streams can generally be characterized by high channel gradients (>2–4%), wide grain-size distributions, channel-spanning bedforms (step-pool systems), shallow flows, common presence of large woody debris and variable channel widths (e.g. Chiari et al. 2010; Nitsche et al. 2011; Zimmermann et al. 2010). These parts of drainage network show extremely rapid rainfall-runoff response (Rickenmann 1997) and present a major flood hazard in a mountainous terrain (Chiari et al. 2010). Nevertheless, bedload transport and hydraulics of steep mountainous streams are poorly understood relative to those of lowland streams (Rickenmann 1997; Aberle & Smart 2003). Supply-limited character together with increased flow resistance are the main reasons for the fact that observed bedload transport in steep torrents may be considerably smaller than that predicted by conventional bedload transport equations (Yager et al. 2007; Chiari & Rickenmann 2011). The introduction of grain and form resistance in order to lower energy gradient helps to correct bedload estimation in high-gradient streams (Rickenmann 2005; Chiari et al. 2010; Chiari & Rickenmann 2011). Some uncertainty in estimation of critical conditions for the beginning of bedload transport in gravel bed streams can also arise due to the significant occurrence of interstitial flows and thus, decrease in potential stream power for sediment transport processes (Carling et al. 2006).

The SETRAC numerical model (Rickenmann et al. 2006), a predecessor of the TOMSED model, was successfully applied in the reconstruction of bedload transport during high-magnitude floods in Alpine catchments (Chiari et al. 2010; Chiari & Rickenmann 2011). TOMSED represents a one-dimensional bedload transport model used to simulate bedload transport at steep slopes developed by Friedl and Chiari (2011) at the University of Natural Resources and Applied Life Sciences, Vienna. Several flow resistance and bedload transport equations have been implanted into the model that enables the use of a specific combination of approaches for particular conditions. Since simple flow resistance and bedload transport equations usually overestimate bedload discharges due to limited sediment-supply conditions and energy loss connected with bed deviations (e.g. steps, pools, and large grains), approaches of form roughness losses need to be taken into account and armouring effects help to calibrate the intensity of bedload transport with field observations. Channel network is defined by nodes, cross-sections and sections. In the TOMSED model, it is also possible to simulate potential erosion or deposition in a stream channel during a flood event and it is possible to set spatial limits of bed elevation changes for individual sections. As stream widths remain constant during the simulation, lateral erosion is not taken into consideration. Input hydrographs can be related to cross-sections as time series in the same way as sediment inputs. For more details about the TOMSED model see Friedl and Chiari (2011), the model is free for download at www.bedload.at.

As we indicated previously, bedload transport estimations in small steep streams represent a challenge for both geomorphologists and water resource engineers due to lack of direct bedload measurements in high-gradient channels. Thus, the main aim of the study was to evaluate the application of the TOMSED v0.1 numerical model in midmountain relief of the Czech landscape. The model was tested on two selected headwater basins of the Western Flysch Carpathians affected by 5/2010 flood. Previously, Galia and Hradecký (2012a, 2012b) investigated critical conditions of the incipient motion of individual boulders moved during range of discharges including

To be quoted as:
Galia, T. – Hradecký, J. (2014): Estimation of bedload transport in headwater streams using a numerical model (Moravskoslezké Beskydy Mts, Czech Republic)
AUC Geographica, 49, No. 1, pp. 21–31

5/2010 event in the study area. Bedload transport was also considered in local gravel-bed rivers with respect to their geomorphic transitions caused by human disturbances (Galia et al. 2012). This time, the authors attempted to evaluate total bedload transport connected with changes in the longitudinal profile, whereas TOMSED simulations were compared to field observations. The authors sincerely hope that the research results will bring a new insight into the knowledge of the behaviour of local headwater streams and improve the management of these forested watersheds.

2. Studied streams

Simulations of bedload transport were conducted in two headwater streams of the Radhošťská hornatina Mts (Moravskoslezské Beskydy Mts), namely Malá Ráztoka and the upper part of Velký Škaredý potok (Table 1, Figure 1, and Figure 2). Lengths of simulated channels were 2100 m of Malá Ráztoka Stream and 1900 m of Velký Škaredý potok respectively, both with 50 m discretisation between individual measured cross-sections.

The streams significantly differ in sediment supply and transport regime (Figure 3). Bedrock outcrops often occur in Malá Ráztoka channel (up to 1/2 of the total length of the longitudinal stream profile) and the stream is recognized as sediment-supply limited. Sediment sources are infrequent bank failures and gullies; non-resistant claystone members of the Godula Formation (lower Cretaceous period) also contribute to sediment delivery in the downstream part of the basin. On the contrary, Velký Škaredý potok lacks bedrock outcrops in the channel bed except for the steepest part of the longitudinal profile consisting in resistant sandstone layers of the middle-part of the Godula Formation. Between 0.0–1.3 km, the stream is located in less-resistant claystones of the Lhoty Member, Veřovice Member and Těšín-Hradiště Formation with a frequent occurrence of bank failures and shallow landslides richly supplying the stream with material. At the bottom of the V-shaped valley of Velký Škaredý potok at 0.0–1.0 km, there is an 8–20 m wide cover of sediments that have been transported there by fluvial processes and also by older debris-flows with related ‘fire hose’ effect (see Šilhán & Pánek 2010). These facts imply that Velký Škaredý potok is rather transport-limited, characterized by relatively unlimited supply conditions, although accelerated vertical erosion was observed in some locations at the bottom of the valley (up to 1 m). Both the basins are also influenced by deep-seated slope deformations in the culmination parts of ridges (e.g. Hradecký & Pánek 2008). Moreover, the relief of the Moravskoslezské Beskydy Mts can be described as polygenetic (Pánek et al. 2009). Spruce and beech forests cover nearly 100% of area

Tab. 1 Characteristics of the studied streams. Symbol “*” indicates different values of sediment parameters for channel-reaches based in generally claystone layers (0.0–1.3 km) with lower values of d_{30}, d_{50} and d_{90}.

	Malá Ráztoka	**Velký Škaredý potok**
Basin area (km²)	2.20	1.06
Altitude (m)	570–1084	510–1129
Channel gradient (m/m)	0.02–0.27	0.03–0.53
Bankfull channel width (m)	3.2–5.1	2.6–4.1
Bankfull hydraulic radius (m)	0.23–0.33	0.22–0.41
d_{30} (mm)	30	30
d_{50} (mm)	55	45*, 55
d_{90} (mm)	220	140*, 180

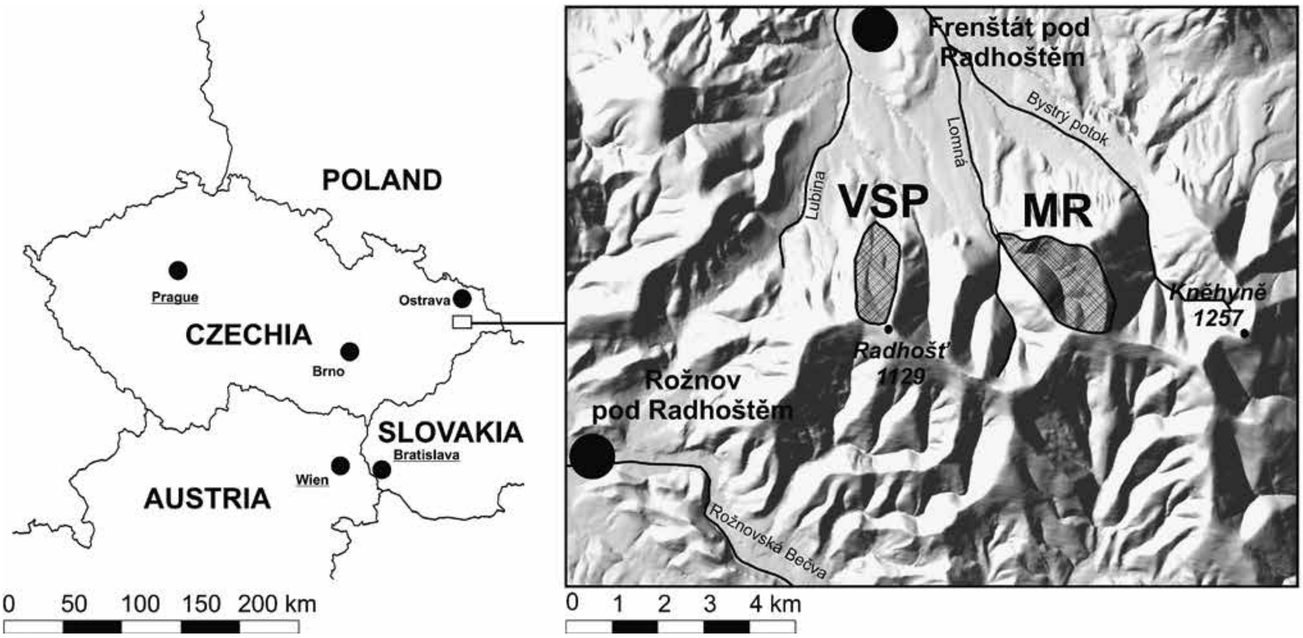


Fig. 1 Location of studied streams: Malá Ráztoka (MR) and Velký Škaredý potok (VSP).

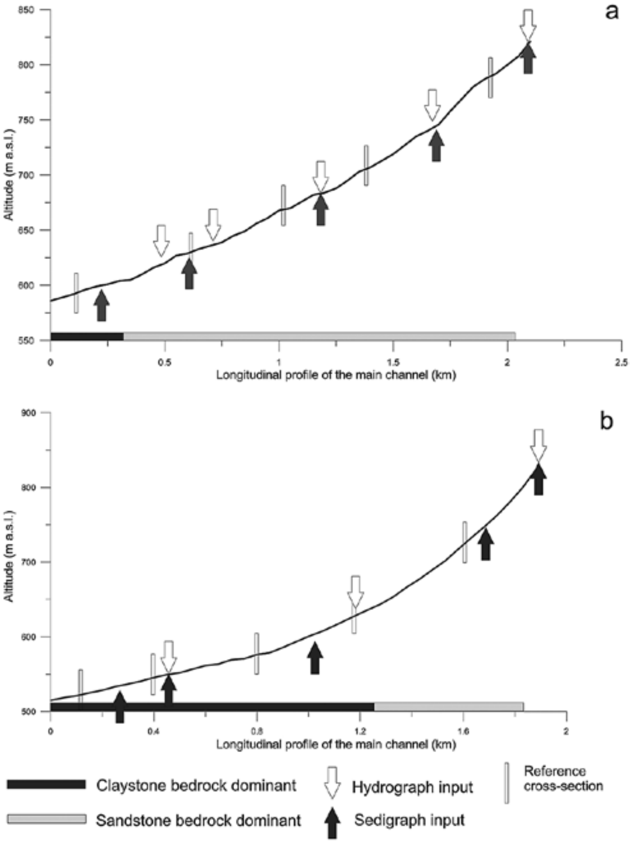


Fig. 2 Schematic longitudinal profiles of Malá Ráztoka (a) and Velký Škaredý potok (b) used in simulations with indicated locations of inputs of water and sediments during a flood event.

of the studied basins. Large woody debris are systematically removed from the local channels by forest management except the uppermost parts of the Velký Škaredý potok basin, where the nature reserve "Radhošť" has been established.

Discharge data from the gauging station in Malá Ráztoka stream, which cover 2.01 km^2 of the basin, were used to reconstruct bedload transport. Local 2-year discharge corresponds to 0.95 m^3 s^{-1}, 10-year discharge to 2.71 m^3 s^{-1}, 25-year discharge to 4.38 m^3 s^{-1} and, finally, 100-year discharge is equal to 8.5 m^3 s^{-1}, as it was derived from the 1954–1993 specific discharge data series (Chlebek & Jařabáč 1995). In addition, mean annual discharge corresponds to 0.061 m^3 s^{-1}, as measured at the gauging station on the stream.

After 5/2010 flood event, erosion and deposition of material was observed in both studied headwater channels. There were some noticeable signs of accelerated vertical erosion in Malá Ráztoka stream at 0.10–0.15, 0.40, 0.50, 0.65–0.75 and 1.9 km from the mouth to Lomná stream. By contrast, depositional processes were evident at 0.05 km, 0.45 km (due to the lowering of bed gradient by one of the check-dams), 1.60–1.65 km and 1.8 km. Velký Škaredý potok indicated some erosional trends at 0.00–0.05 km, 0.90–0.95 km, 1.20–1.25 km (due to the action of a sluice as a barrier) and 1.65 km upstream from the ford and the reservoir near the first building of Trojanovice village, which is the end of the studied channel. As it was mentioned above, almost the whole profile of the longitudinal profile of the stream is characterized by a sufficient amount of sediments, whereas most of the deposits were reactivated and transported during 5/2010 flood. In both headwater streams, large woody debris also contributed to create alluvial accumulations in active channels. Three check dams and a measuring flume have been constructed in the bottom part of Malá Ráztoka longitudinal profile between 0.25 and 0.45 km during the second half of 20th century. They naturally affect the transport dynamics during flood events. Similarly, a sluice built across the road at 1.25 km of Velký Škaredý potok stream has recently been acting as a barrier since sediments clogged up inside. For bedload material, we use a bulk density value of 2550 kg m^{-3}, which is valid for sandstones of the Godula Member.

3. Methods

During the summer season of 2011, channel cross-sections were measured using tape in 50-m-long intervals. Laser rangefinder including a clinometer was used to obtain individual vertical distances between

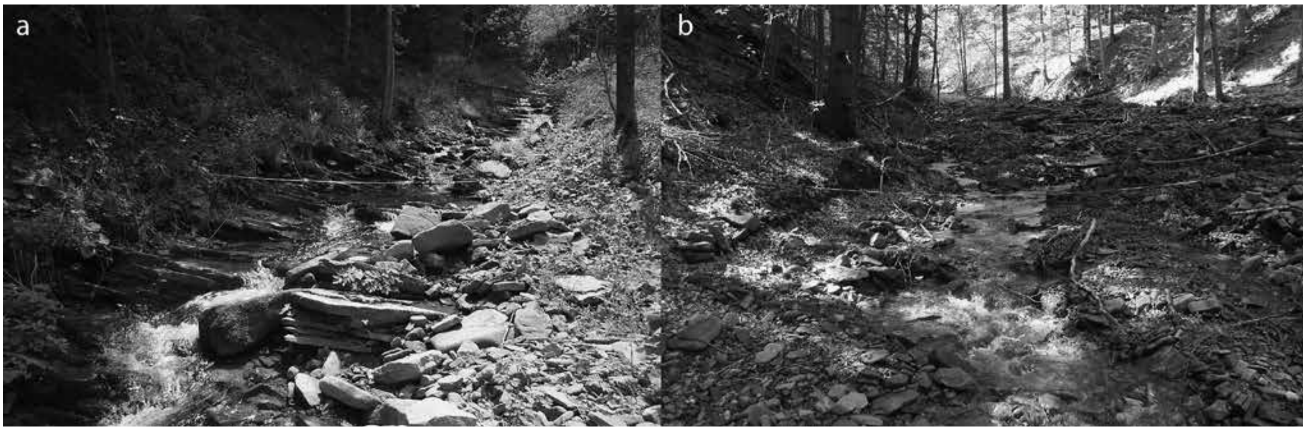


Fig. 3 Supply-limited reach with frequent bedrock outcrops in Malá Ráztoka stream (a) and well sediment-supplied (transport-limited) reach of Velký Škaredý potok stream (b).

individual cross-sections. One type of sediment-size distribution was applied into the Malá Ráztoka modelled profile due to relatively uniform lithology of channel-bed material (sandstone layers of the Godula Member), whereas in the case of Velký Škaredý potok, two diverse types of sediment-size distribution were applied for cross-sections based in sandstone layers and claystone-prevailing formations. Measurement of the sediment size was conducted by the Bevenger-King sampling method (1995), which had been derived from the Wolman method (1954).

Based on the data series measured at Malá Ráztoka gauging station, a 6-hour interval of time steps between 16 May 2010 0:00 and 20 May 2010 0:00 was used to simulate bedload transport during an event using the TOMSED v0.1 model. The flood culminated on 17 May 2010 at 12:00 with 3.96 $m^3 s^{-1}$ discharge, which corresponds to a 20–25y flood, as based on published specific discharge data series (1954–1993) from the basin (Chlebek & Jařabáč 1995). Then approximate inputs of hydrographs were added to the stream profile (Figure 2). These inputs were dependent on their sources and sub-basin areas and their total sum was equivalent to the final observed discharge at the gauging station. Discharges for Velký Škaredý potok stream were derived from Malá Ráztoka gauging station using a simple specific discharge method. Consequently, individual inputs for the stream profile were developed. This showed that the flood culmination in the modelled outlet of Velký Škaredý potok (1.06 km^2, 2.0 km length) reached values of 2 $m^3 s^{-1}$. Some inputs of sediments ('sedigraphs') were also assigned to locations in both stream profiles where the delivery of material is expected during a flood event (large active bank failures, outfalls of gullies). However, no exact measuring was conducted of material delivery to the fluvial system of the study area; therefore, quantitative values of inputs were rather adjusted to modelled bedload discharges and geomorphic evidence in the field.

In order to make an exact comparison of results, identical flow resistance and bedload equations were used as those presented in the paper of Chiari and Rickenmann (2011). Flow velocity v was computed by well-established Manning equation including hydraulic radius of channel (R), energy gradient (S) and total flow resistance represented by Manning coefficient n_{tot}

$$v = R^{0.67} S^{0.5} / n_{tot}, \qquad (1)$$

although some authors (e.g. Ferguson 2010) suggest the application of more appropriate flow-resistance approaches for shallow flows and steep gradients. TOMSED also allows the use of logarithmic and power approaches, but we only investigated very small differences comparing the results obtained by Manning equation without reduced energy gradient, logarithmic law after Smart and Jäggi (1983) and power law after Rickenmann et al. (2006). To obtain total flow resistance during a known discharge Q in steep channels >0.008 m/m, Rickenmann (1996) derived the equation:

$$1/n_{tot} = (0.97 g^{0.41} Q^{0.19}) / (S^{0.19} d_{90}^{0.64}), \qquad (2)$$

where g is the acceleration of gravity and d_{90} is the 90th percentile of grain-size cumulative curve. Due to the importance of bedforms in steep channels, Rickenmann (2005) proposed a general empirical function of increased form resistance which is based on dividing total resistance into grain and form resistance and subsequent determination of reduced energy gradient S_{red} related to grain resistance n_r only. The equations take forms:

$$n_r / n_{tot} = 0.092 S^{-0.35} (D / d_{90})^{0.33} \qquad (3)$$

and

$$S_{red} = S (n_r / n_{tot})^a, \qquad (4)$$

where D means flow depth and a is an exponent occupying the range $1 \le a \le 2$. Chiari and Rickenmann (2011) used a equal to 1.0 and 1.5 to correct bedload estimations according to field observations in Alpine streams. In order to compute unit bedload transport q_b, Rickenmann's (2001) equation was used:

$$q_b = 3.1 (d_{90} / d_{30})^{0.2} (q - q_c) S^{1.5} (s - 1)^{-1.5}, \qquad (5)$$

where d_{30} is the 30th percentile of a grain-size cumulative curve, q is unit discharge, q_c means critical unit discharge for the incipient motion of bed material and s is a ratio of sediment density and fluid density ($ς_s/ς$). Finally, critical unit discharge q_c necessary to initiate bedload transport may be obtained by the equation originally proposed by Bathurst et al. (1987):

$$q_c = 0.045 (s - 1)^{1.67} g^{0.5} d_{50}^{1.5} S^{-1.12}, \qquad (6)$$

where d_{50} represents median grain size of the bed surface layer. Later Rickenmann (1990) modified the value 0.045 in Eq. (6) to the value 0.065, which can be understood as the substitution of dimensionless critical shear stress. Dimensionless shear stress occurs commonly in sediment transport equations. TOMSED does not allow including specific values of dimensionless critical shear stress for individual cross-sections or channel reaches. This fact affects the intensity of bedload transport particularly in steeper reaches (further in the text) via some other approaches (e.g. logarithmic flow resistance) implanted in the model. Although the simulation of an armour layer is possible in the TOMSED model, no such layer was supposed in Velký Škaredý potok stream due to high sediment delivery into headwater segments. At the same time, frequent occurrence of bedrock in Malá Ráztoka stream prevents the formation of a continuous armour layer, therefore this criterion was neglected the in simulation of both streams.

4. Results

4.1 Malá Ráztoka stream

The simulation of 5/2010 flood event was performed by means of Eq. (1–6) for four different stages: *i*) no sediment feed, no energy slope reduction and no limit for vertical erosion was introduced (10 m limit was set for all cross-sections), *ii*) rough limits were defined for erosion depth in accordance with field observations: 0.2 m for bedrock-cascade reaches, 1.0 m for all the other reaches), *iii*) sediment feed was introduced in locations with supposed delivery of material during the flood, and finally *iv*) slope reduction was established for the simulated stream, while parameter *a* was set to 1.5 in Eq. (4). As Figure 4 illustrates, stage *i*) shows unrealistically large erosion (up to 10 m limit) in the upper part of the basin. After introducing incision limits (stage *ii*), erosive trends prevailed in the whole longitudinal profile, except the downstream part near the junction with Lomná stream. This roughly corresponds with field observations according to which the channel of Malá Ráztoka is generally considered as a sediment supply limited system. The supply of sediment feed (*iii*) only created small positive peaks in places of supposed sediment supply due to installed sedigraphs. The overall character remained to be erosional. Reducing the slope (stage *iv*), several positive peaks were computed where deposition might occur, although the trend of erosional processes was preserved along the majority of the stream length. Depositions simulated at 0.4 km, 1.7 km and 1.9 km roughly correspond to after-flood field observations. In the most downstream part, by the junction with Lomná stream, material accumulation occurs both in the model and in the field. On the contrary, the TOMSED model predicts depositional processes for the zones at 0.6–0.75 km, 1.05 km and 1.45–1.55 km, but most of the reaches are recently eroded bedrock-cascades with no signs of aggradation.

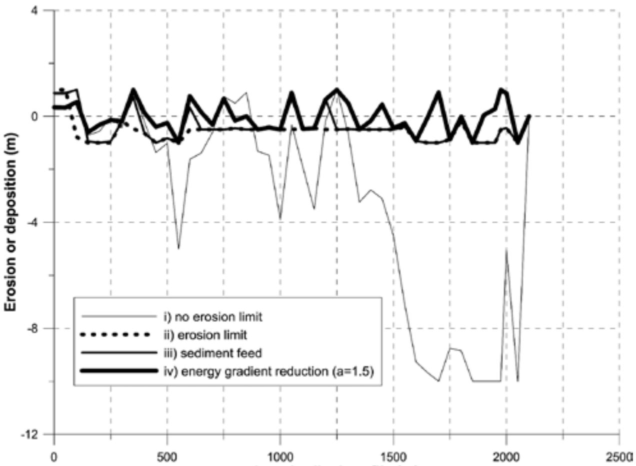


Fig. 4 Modelled changes in the longitudinal profile due to erosion or deposition during the 5/2010 flood in Malá Ráztoka stream.

Figure 5 displays the potential intensity of bedload transport at five selected cross-sections if considering stage *iv*) (limits in erosion depth, introduced sediment feed and slope reduction with $a = 1.5$). The simulation supposed the highest bedload transport at 0.1 km, 0.6 km and 1.4 km cross-sections, where the values of up to 40 m^3/h (100t/h) of transported material were reached during the peak discharge. On the contrary, the smallest intensity was predicted in the uppermost investigated reach (1.8 km), which could correspond to the real situation during the 5/2010 flood event, according to smaller field observed dynamics of erosional or depositional processes after that event. Flood situation and damages of the afternoon of 18th May (about 24 hours after flood culmination) were evaluated in the downstream part of the basin (0.0–0.5 km). Bank erosion as well as some fresh material deposits, mainly upstream of the check-dams, were noticed with a higher frequency and of a greater size than those observed in the uppermost reaches. This implies that the TOMSED model can predict well the intensity of bedload transport, which facilitates the comparison of individual cross-sections and reaches. Absolute values of computed transported material sound realistic, but no exact measuring of bedload transport intensity has been carried out in local high-gradient streams to calibrate modelled calculations.

Total bedload transport during the 5/2010 flood event was estimated in a range of 380–860 m^3 (970–2200 t) of transported coarse material in selected cross-sections at 0.1 km, 0.6 km, 1.0 km, 1.4 km and 1.8 km (Figure 6).

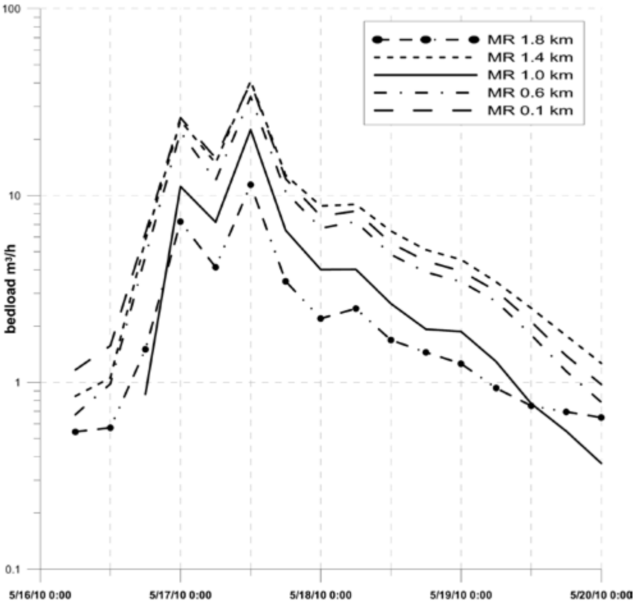
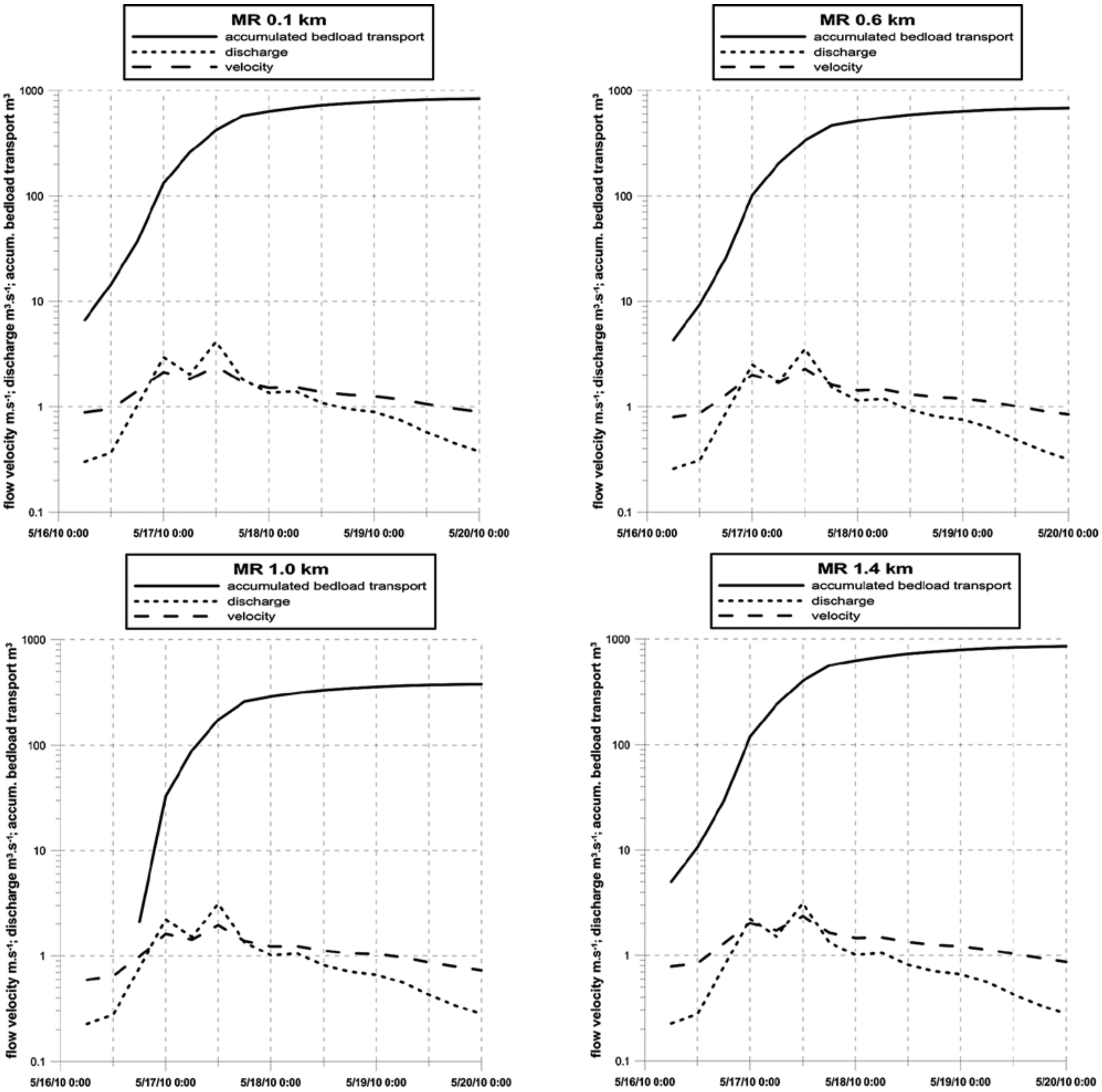


Fig. 5 Intensity of bedload transport during the 5/2010 flood at 0.1 km, 0.6 km, 1.0 km, 1.4 km and 1.8 km of the longitudinal profile of Malá Ráztoka.



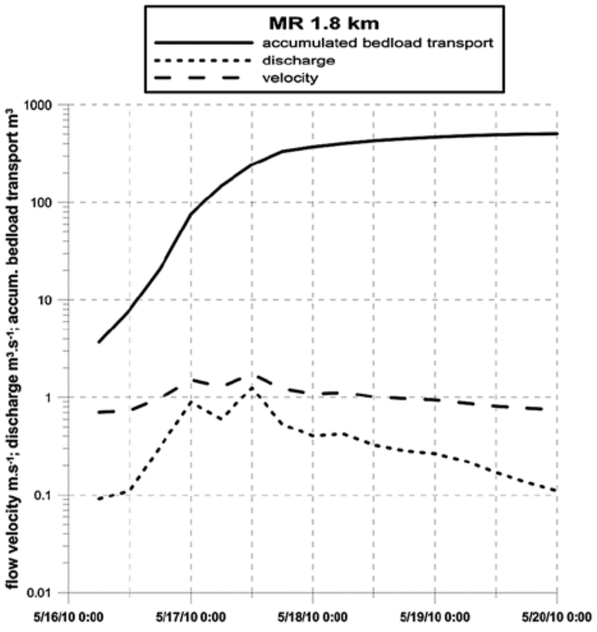


Fig. 6 Flow velocity, discharge and cumulative bedload transport simulated at the reference cross-sections of Malá Ráztoka stream during the 5/2010 flood.

The cross-sections at 0.1 km together with 1.4 km indicated the highest amounts of transported material with values exceeding 800 m^3 (2000 t). It is notable that suspended transport is not considered in the reported total volumes of transported sediment. Mean modelled flow velocities were equal to 1.7–2.4 m s^{-1} during the flood peak when the highest values were rather observed in the downstream reaches (0.1 km and 0.6 km) and 1.4 km cross-section.

4.2 Velký Škaredý potok stream

Flood event discharge data were computed using a simple specific discharge method for the modelled outlet of Velký Škaredý potok stream. Consequently, individual inputs of hydrographs and sedigraphs were made for particular reaches based on local assumptions. Similarly to simulations in Malá Ráztoka, four different stages were used for bedload modelling. In relation to high amounts of potentially mobilized material at the bottom of the stream valley, fairly high dynamics of erosional and accumulation processes was observed in the lower parts (0.0–1.0 km) after 5/2010 flood. Resulting intensity of bedload transport in lower parts (0.1, 0.4 and 0.8 km) at the stage *iv*) was lower than the intensity calculated for the 0.1, 0.6 and 1.4 km of Malá Ráztoka stream and even lower than the intensity obtained in the uppermost cross-sections of Velký Škaredý potok (e.g. 1.6 km). This did not correspond to the real situation, especially to total sediment availability for bedload transport and observed erosion and accumulation forms after the flood in both the streams. To increase simulated bedload discharge in lower parts of Velký Škaredý potok, the exponent *a* in Eq. (4) was optimized to value 1.0 and included in the stage *v*).

Stage *i*) simulation showed as huge incision in the uppermost part of the longitudinal profile during the flood as in Malá Ráztoka, but some deposition was also observed in the lowest parts (Figure 7). The stages void of slope reduction and erosion limits (stage *ii*) and stage *iii*)) resulted in a similar model of the longitudinal profile, although some positive peaks occurred due to inputs of sediment in stage *iii*). Both *ii*) and *iii*) stages indicated the erosional trend along almost the whole length of the simulated channel even though little aggradation was found within stage *iii*) at 0.15–0.40 km. Stage *v*) (*a* = 1.0) simulation showed higher amplitudes in erosional/depositional trends than stage *iv*) (*a* = 1.5), but the courses of the trends were similar. Highly fluctuating trends with noticeable peaks of erosion and deposition were observed in the upper parts of the stream, whereas a relative equilibrium between incision and aggradation was observed further downstream. Compared to the real situation, no larger erosional or depositional forms were observed in the upper part of the profile, except at the 1.6 km, which was recognized well by the simulation. Therefore, the use of *a* = 1.5 in Eq. (4) would be appropriate for upstream reaches. On the contrary, as it was mentioned above, higher intensity of bedload transport had been assumed in the lower parts (0.0–1.0 km) and thus *a* exponent equal to 1.0 in Eq. (4) led to better agreement with the real situation in the downstream parts. Also, trends of incision at 0.00–0.05 km and 0.90–0.95 km and downstream the jammed sluice at 1.20–1.25 km were exposed well by the simulations within *iv*) and *v*) stages. Deposition of material up the sluice (1.25 km) was indicated in the modelled profiles as well.

Reference cross-sections of 0.1 and 0.4 km showed higher bedload discharges during the flood event (Figure 8) with peak values of about 50 m^3/h (125 t/h) for *a* = 1.0 at Eq. (4). Bedload transport was of a higher intensity in these lower parts due to large potential sediment delivery to the stream. The main channel of the stream

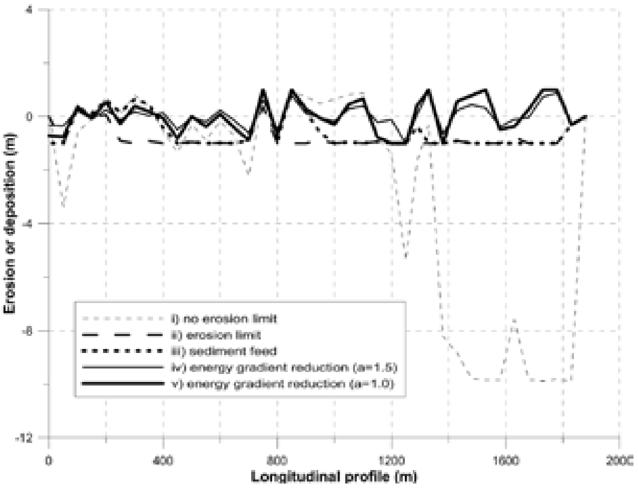


Fig. 7 Modelled changes in the longitudinal profile due to erosion or deposition during the 5/2010 flood in Velký Škaredý potok stream.

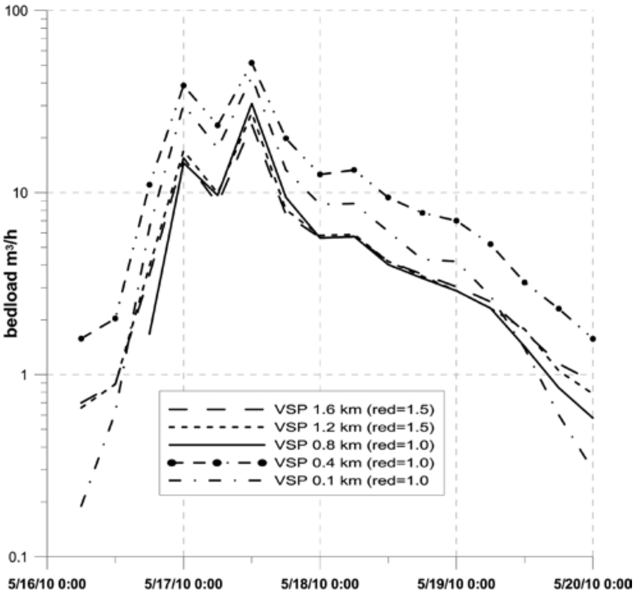


Fig. 8 Intensity of bedload transport during the 5/2010 flood at 0.1, 0.4, 0.8, 1.2 and 1.6 km of the longitudinal profile of Velký Škaredý potok.

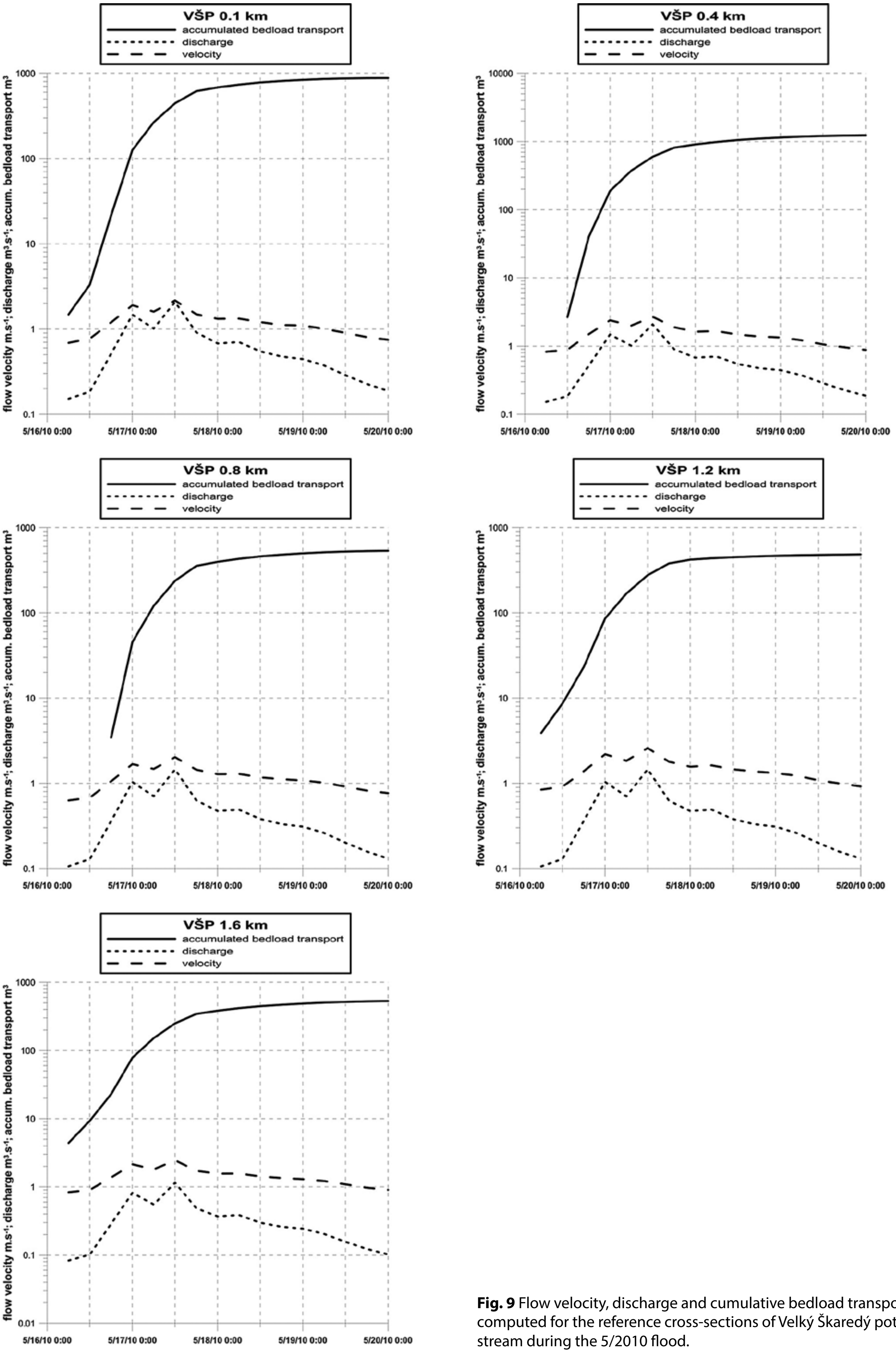


Fig. 9 Flow velocity, discharge and cumulative bedload transport computed for the reference cross-sections of Velký Škaredý potok stream during the 5/2010 flood.

changed its course in the relatively wide valley bottom (up to 20 m) during the flood event as transported material was stored or eroded by the stream flow. Bedload transport reached values of up to 25–30 m^3/h (75 t/h) for $a = 1.5$ at Eq. (4) in the upper part of the stream during the flood culmination. With respect to local channel gradients of 0.20 m/m, some movement of the type of smaller debris flows might have been reactivated in the channel.

Total volumes of transported bedload are a bit higher than those simulated for Malá Ráztoka. Values were computed in a range of 480–1240 m^3 (1220–3170 t) for 0.1, 0.4, 0.8, 1.2 and 1.6 km channel cross-section (Figure 9). Smaller moved amounts were assumed in connection with the upper part of the stream (0.8, 1.2 and 1.6 km) with transport rates of 480–540 m^3, whereas the downstream part (0.1 and 0.4 km) indicated much higher intensity with values of 880 m^3 and 1240 m^3 during the flood event. It was in agreement with the real situation, when a higher activity of depositional and erosional processes was recognized in the lower part of the basin after the flood event. The highest mean flow velocity during the peak discharge was simulated in the uppermost part of the longitudinal profile (2.5 m s^{-1}) in accordance with the highest channel gradient (up to 0.5 m/m) at 0.4 km (2.7 m s^{-1}). On the other hand, the lowest velocities of about 2.0 m s^{-1} were only computed for 0.1 and 0.8 km cross-sections.

5. Discussion

This study makes use of the same approaches of bedload transport simulation that were originally applied in the reconstruction of bedload transport in Alpine streams with larger basin areas (6–180 km^2) and much higher peak discharges (24–140 m^3 s^{-1}) by Chiari and Rickenmann (2011). They regarded the application of macro-roughness (Eq. 3 and Eq. 4) with the calibrated parameter *a* in Eq. (4) in a range of 1.0–1.5 as the best-fitting predictor of bedload transport in such high-gradient streams, although they admitted that the quantification of macro-roughness remained difficult. According to our investigations, *a* exponent equal to 1.5 may fit to reaches with a limited supply of sediments for further transport and occurrence of well-developed bedforms (interlocked structures of individual steps). On the other hand, a lower value (1.0) most likely corresponds to parts of the longitudinal profile with almost unlimited delivery of material and thus less developed or infrequent step structures or other flow resistance bedforms. The degree of stability of steps and individual large grains affect potential bedload transport due to kinetic energy losses. Moreover, frequent bedrock outcrops in Malá Ráztoka stream substitute alluvial steps in flow energy dissipation and as a result they permanently act as resistance forms until they are buried by sediments during rather rare aggradation.

Another problem arises in connection with the incipient motion of grain or bedload layer according to Eq. (6). Chiari and Rickenmann (2011) used a single value 0.065 after Rickenmann (1991), whereas Bathurst (1987) originally recommended the value 0.045. This value can simply be understood as an expression of dimensionless shear stress or the so-called 'Shields parameter'. The TOMSED v0.1 model only allows the application of a single value of Shields parameter for a simulated stream, but it is possible the application of various values in the newest 0.2 version. Lamb et al. (2008) and later also Parker et al. (2011) demonstrated the dependence of dimensionless shear stress and at the same time the stability of individual grain on the channel gradient, pointing on a positive trend between increasing channel gradient and dimensionless shear stress. The specification of dimensionless shear stress values for various ranges of channel gradients will possibly make the resulting intensity of bedload transport more accurate for one-grain size models, especially in case of steeper gradients. Lenzi et al. (2006) also assumed different values of dimensionless shear stress, this time for different sizes of bed particles. His method is based on the comparison of a certain grain diameter with bed surface d_{50} or d_{90}. Applying this approach on Malá Ráztoka stream, the calculated values of dimensionless shear stresses were from 0.02 for the 0.5 m boulders to low values of about 0.400 for 0.01 m grain diameters (Galia & Hradecký, 2011). This is further applicable to the calculations of fractional bedload transport, which the TOMSED model allows as well.

Exact values of real bedload transport discharges are not known for the Beskydian basins for any flood event, which makes the validating of the TOMSED model complicated for local mid-mountain conditions. Alpine torrents investigated by Chiari and Rickenmann (2011) indicated much higher intensity of bedload transport with total volumes of transported sediments commonly exceeding 10,000 m^3 during flood events due to larger basin areas and peak discharges than those simulated in this paper (maximum 1240 m^3 at the 0.4 km cross-section in Velký Škaredý potok). Kabina and Halaj (2004) reported storage of 3310 m^3 of bedload material during an extraordinary 100y flood event in Slovakian mid-mountain stream Hrondín (basin area 7.25 km^2, peak 100y discharge 27.8 m^3 s^{-1}). By contrast, Lenzi et al. (1999) observed peak bedload discharges of about 250 m^3 h^{-1} only resulting in 900 m^3 of accumulated bedload transport in the sediment supply limited Alpine basin of Rio Cordon (5 km^2) during a 30–50y discharge. As for smaller events (1–5y discharge), the bedload rate did not exceed the values of 6 m^3 h^{-1} in Rio Cordon. On the other hand, TOMSED relatively well estimated particular reaches with prevailing erosional or depositional processes, mainly in Velký Škaredý potok headwater stream, although the simulations were conducted on after-flood measured cross-sections.

6. Conclusions

The TOMSED model, originally developed for Alpine torrents, was applied first time in the conditions of the Czech midmountains. The Beskydian headwater streams of Malá Ráztoka and Velký Škaredý potok were selected in order to simulate the 5/2010 flood (20–25y discharge). Maximum intensity of bedload transport in a range of 10–50 m^3/h (25–125 t/h) was computed during the flood event peak discharge and reaches with potential erosion and deposition were estimated. Although the inexistence of field measurement in the studied locality made it impossible to validate simulated values of bedload sediment transport, the TOMSED model indicated the reaches with recent incision or aggradation acceptably well. The approach of reducing the energy gradient for bedload transport was considered to be reliable for the estimations of coarse sediment transport and for the determination of the intensity of incision or aggradation in accordance with the resulting simulated longitudinal profiles. Decrease in flow resistance partitioning was suggested for high sediment supply related with less developed bedforms. The paper also discussed possibilities related to accuracy estimations of bedload transport by introducing individual values of dimensionless critical shear stress for different channel gradients or size fractions of transported material.

Moreover, obtained calculations may improve the management of local high-gradient streams (stream restorations, protections against flood damages etc.) due to bringing relatively exact values of the maximal intensity of bedload transport and volumes of totally transported material during high-magnitude flood event. Our results indicate that the TOMSED model is applicable in high-gradient streams of the Czech landscape, nevertheless the further improvement of simulated bedload transport is naturally related to some exact measurement in any of small mountain stream.

Acknowledgements

The research was supported by the Student Competition Grant No. SGS6/PřF/2011. We would like to thank the Forestry and Game Management Research Institute, Frýdek-Místek branch, for providing valuable discharge data from the Malá Ráztoka instrumental basin. 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, 259–269.

BATHURST, J., GRAF, W., CAO, H. (1987): Bed load discharge equations for steep mountain rivers. In Thorne, C. R., Bathurst, J. C., Hey, R. D. (eds): Sediment Transport in Gravel-Bed Rivers. Wiley, New York, 453–477.

BEVENGER, G. S., KING, R. M. (1995): A pebble count procedure for assessing watershed cumulative effects. USDA Forest Service Research Paper RM-RP319 Fort Collins, CO: U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station.

CARLING, P. A., WHITCOMBE, L., BENSON, I. A., HANKIN, B. G., RADECKI-PAWLIK, A. M. (2006): A new method to determine interstitial flow patterns in flume studies of sub-aqueous gravel bedforms such as fish nests. River Res. Applic. 22, 691–701.

CHIARI, M., FRIEDEL, K., RICKENMANN, D. (2010): A one-dimensional bedload transport model for steep slopes. Journal of Hydraulic Research 48, 2, 152–160.

CHIARI, M., RICKENMANN, D. (2011): Back-calculation of bedload transport in steep channels with a numerical model. Earth Surface Processes and Landforms 36, 805–815.

CHLEBEK, A., JAŘABÁČ, M. (1995): Posouzení závislosti přírůstku lesních porostů v Beskydech na klimatu metodou faktorové analýzy. Lesnictví 41, 9, 426–431.

FERGUSON, R. (2010): Time to abandon the Manning equation? Earth Surface Processes and Landforms 35, 1873–1876.

FRIEDL, K., CHIARI, M. (2011): A one-dimensional transport model for steep slopes. Manual Tomsed version 0.1 beta. Online at http://www.bedload.at.

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, 238–250.

GALIA, T., HRADECKÝ, J. (2012a): Critical conditions for the beginning of coarse sediment transport in the torrents of the Moravskoslezské Beskydy Mts (Western Carpathians). Carpathian Journal of Earth and Environmental Sciences 7, 4, 5–14.

GALIA, T., HRADECKÝ, J. (2012b): Bedload transport and flow resistance in steep channels – introduction to the problem in context of mountain basins of Central European region. Acta Universitatis Carolinae Geographica 47, 1, 23–33.

GALIA, T., ŠKARPICH, V., HRADECKÝ, J. (2012): Dnový transport sedimentů v souvislosti s transformací geomorfologického režimu štěrkonosných toků Moravskoslezských Beskyd. Geografie, 117, 1, 95–109.

HRADECKÝ, J., PÁNEK, T. (2008): Deep-seated gravitational slope deformations and their influence on consequent mass movements (case studies from the highest part of the Czech Carpathians). Natural Hazards 45, 235–253.

KABINA, P., HALAJ, P. (2004): Zhodnotenie vplyvu prehrádzky na zmeny na toku Hrondín. In Šiška, B., Igaz, D. (eds): Bioklimatologické pracovné dni 2004, 23.–26. 8. 2004, Viničky, okr. Trebišov, Slovensko.

LAMB, M., DIETRICH, W., VENDITTI, J. (2008): Is the critical shear stress of incident sediment motion dependent on channel-bed slope? J. Geophys. Res. – Earth Surface 113. F02008, doi:10.1029/2007JF000831.

LENZI, M. A., D'AGOSTINO, V., BILLI, P. (1999): Bedload transport in the instrumented catchment of the Rio Cordon, Part I: Analysis of bedload records, conditions and threshold of bedload entrainment. Catena 36, 171–190.

LENZI, M. A., MAO, L., COMITI, F. (2006): When does bedload transport begin in steep boulder-bed stream? Hydrologic. Process. 20, 3517–3533.

NITSCHE, M., RICKENMANN, D., TUROWSKI, J. M., BADOUX, A., KIRCHNER, J. W. (2011): Evaluation of bedload transport predictions using flow resistance equations to account for macro-roughness in steep mountain streams. Water Resources Research, 47, W08513, doi:10.1029/2011WR010645.

PÁNEK, T., HRADECKÝ, J., ŠILHÁN, K. (2009): Geomorphic evidence of catastrophic flow type landslide in the mid-mountain ridges of the Western Flysch Carpathian Mountains (Czech Republic). International Journal of Sediment Research 24, 88–98.

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, 51–65.

RICKENMANN, D. (1990): Bedload transport capacity of slurry flows at steep slopes. Mitteilung 103 der Versuchsanstalt für Wasserbau, Hydrologie Glaziologie, ETH Zürich.

RICKENMANN, D. (1996): Fliessgeschwindigkeit in Wildbächen und Gebirgsflüssen. Wasser Energie Luft 88, 298–304.

RICKENMANN, D. (1997): Sediment transport in Swiss torrents. Earth Surface Processes and Landforms 22, 937–951.

RICKENMANN, D. (2001): Comparison of bed load transport in torrents and gravel bed streams. Water Resources Research 37, 3295–3305.

RICKENMANN, D. (2005): Geschiebetransport bei steilen Gefällen. Mitteilung 190 der Versuchsanstalt für Wasserbau, Hydrologie und Glaziologie, ETH Zurich, p. 107–119.

RICKENMANN, D., CHIARI, M., FRIEDL, K. (2006): SETRAC: A sediment routing model for steep torrent channels. In Ferreira, R., Leal, E. A. J., Cardoso, A. (eds): River Flow 2006, Vol. 1, Taylor & Francis, London, 843–852.

SMART, G, JÄGGI, M. (1983): Sedimenttransport in steilen Gerinnen. Mitteilung 64 der Versuchsanstalt für Wasserbau, Hydrologie und Glaziologie, ETH Zurich.

ŠILHÁN, K., PÁNEK, T. (2010): Fossil and recent debris flows in medium-high mountains (Moravskoslezské Beskydy Mts, Czech Republic). Geomorphology 124, 238–249.

WOLMAN, M. G. (1954): A method of sampling coarse river-bed material. Transactions American Geophysical Union 35, 6, 951–956.

YAGER, E. M., KIRCHNER, J. W., DIETRICH, W. E. (2007): Calculating bed load transport in steep boulder bed channels. Water Resources Research 43: W07418.

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.

RESUMÉ

Použití numerického modelu pro odhad dnového transport sedimentů v pramenných tocích (Moravskoslezské Beskydy, Česká republika)

Numerický model TOMSED, původně vyvinutý pro alpské bystřiny, byl úspěšně aplikován na povodňovou událost 5/2010 v pramenných tocích české části flyšových Západních Karpat. Ve výpočtech byl použit Manningův vztah s rozděleným drsnostním koeficientem na drsnost forem a drsnost zrn společně s redukovaným energetickým gradientem a limitní hloubkou potenciální eroze. Chybějící přímé měření transportu dnových sedimentů komplikovalo validaci modelu pro místní pramenné toky, nicméně korytové úseky se současnými erozními či akumulačními trendy model určil poměrně přesně. V toku bohatě dotovaném sedimenty byly vypočteny hodnoty dnového transportu sedimentů v rozmezí 480–1240 m^3, tok s limitovanou donáškou sedimentů vykázal hodnoty tohoto parametru v rozmezí 380–860 m^3. Bylo zjištěno, že korytové úseky s vyšší dodávkou sedimentů mají méně vyvinuty dnové formace (např. stupně) a tak u takových úseků vede nižší celková korytová drsnost k vyšší intenzitě dnového transportu 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: tomas.galia@osu.cz, jan.hradecky@osu.cz