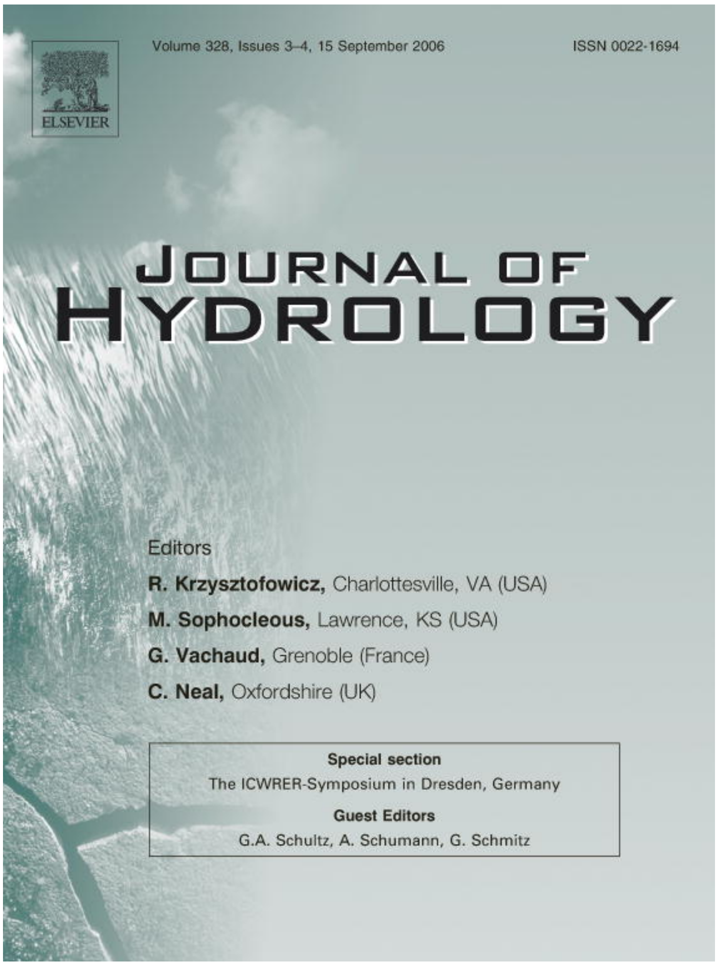
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Flow pattern and sediment transport in a braided river: The ''torrent de St Pierre'' (French Alps)

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Summary In order to bring some understanding on the mountain stream dynamics, we report measurements of flow and sediment transport leveled in a proglacial gravel-bed river, the ''torrent de St Pierre'' in the French Alps. The river exhibits a modest discharge (<5 m^3/s) during most of the glacier melting season. A braiding plain developed thanks to the occurrence of a massive landslide, providing a small and simplified study area quite similar to recent experimental studies. A mass balance was established at the outlet of the braiding plain. Our measurements indicate that, in the range of flows measured, the dominating transport mode is suspension. Though less important, bed load transport is far from being negligible. Dissolved load eventually appears to be very small compared to solid transport. Analysis of velocity profile measurements shows that in this highly turbulent and shallow stream, the use of a logarithmic form fails to recover the velocity profile and to estimate the shear velocity of the flow. A uniform Chézy-like relationship is shown to be valid for the velocity with friction a coefficient considered as constant over the range of our measurements. Accordingly, the only relevant velocity for transport description is the average velocity. Coupled measurements of bed load and velocity show that bed load transport is related to the average velocity through a power law. Eventually, a correlation between bed load transport and suspended load transport is evidenced and discussed.

Introduction

River sediment transport can be divided in three modes: solutions resulting from chemical erosion, and suspension and bed load resulting from mechanical erosion of rocks. In mountain streams, solid load is often assumed to prevail but few mass budgets, integrating these three modes, are available to confirm this assumption. If high concentrations of suspended particles have been observed in glacier melting controlled rivers (Bogen, 1980, 1995), few good data sets of bed load transport exist for these streams (Wohl,

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2000). Among them, one must cite the data base of Williams and Rosgen (1989) which gathers measurements performed on 93 American streams and the work of Ryan et al. (2005). Numerous works, based on experiments, provide descriptions and models for bed load transport. However, field measurements, when available, produce data, which are not always in good agreement with experimental findings. The works of Carson and Griffiths (1987) and Gomez and Church (1989) based on transport equations comparison, are not only in disagreement, but claim that the success rate of many formulas is low (Wohl, 2000). Among the multiple causes of misestimation, the evaluation of the shear stress, commonly employed in the transport laws, is based on strong approximations on the form of velocity profiles, such as the applicability of the Karman–Prandtl logarithmic form for instance (Yalin, 1972). In the field, such velocity profiles may vary substantially both in space and time. This problem is particularly acute for gravel bed braided streams, where turbulence and roughness are high, and flow depth is low. The presence of boulders in the bed modifies the flow considerably and tends to produce s-shaped, linear, or random-shaped velocity profiles (Marchand et al., 1984; Bathurst, 1988; Jarrett, 1991; Byrd, 1997; Wohl, 2000). In this case, the use of log or semi-log velocity profiles may result in a significant overestimation of the flow resistance (Bathurst, 1994; Wohl, 2000). Considering that turbulence can be related to the mean velocity and to the roughness k_s, the use of mean velocity as the relevant velocity for transport relations, might lead to significant results. However, this cannot be done without collecting a large number of data in order to average over the turbulent fluctuations of velocity.

This paper presents a range of field measurements performed on a highly dynamic proglacial gravel-bed river in the French Alps. Because of the presence of a landslide that filled the upper valley, the river exhibits a braiding pattern over a modest plain formed by the accumulation of sediments at the outlet of the glaciers. This braiding plain, unusually high in altitude, provides a simplified study area which is quite similar to our recent experimental studies (Métivier and Meunier, 2003; Meunier and Métivier, 2006). The transport processes presented in this paper are then limited to the case of a confined braided steam with a modest low-flow glacier runoff. The three transport modes have been measured in order to establish a full mass balance and a comparative study of their respective contribution. We use both shear velocity and mean velocity and discuss their relevance for transport law computations. A rapid presentation of the field site and of the measurement protocol is exposed in the next section. Then we then present the results on mass balance, river hydraulics, bed load and suspended load transport and discuss the results.

Field site and methods

River description

Our field site is the torrent de St Pierre, in the massif des Ecrins, one of the highest in the French Alps. The torrent de St Pierre has its source at the confluence of three torrents, coming from the ''Glacier Blanc'', the ''Glacier Noir'' and the ''Glacier de la Momie'' (Fig. 1, left). The site is a specific alluvial plain, called ''Pré de Madame Carle'', which spreads over one and a half kilometers along the river, and is about 500 m in its maximum width. This alluvial plain was created by a landslide which filled up the valley and forced sedimentation of eroded material from the glaciers (Fig. 1, right). The slope of this plain is significantly lower than the slope of the feeding glacial outwash. A GPS topographic profile of the alluvial plain has been done from the outlet of the ''pré de Madame Carle'' to the upstream torrent of the ''Glacier Noir'' (Fig. 2). The average plain gradient is about 2.5%. Bed material is mainly of gravel size

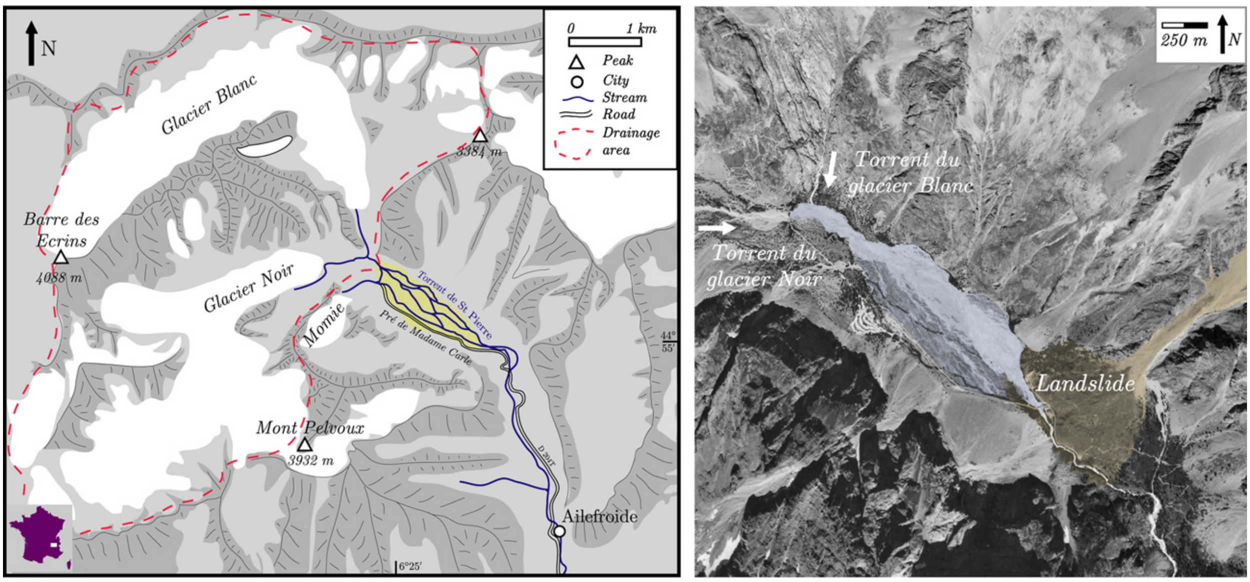

Figure 1 The torrent de St Pierre is fed by the melting of three glaciers: the ''Glacier Blanc'', the ''Glacier Noir'' and the small ''Glacier de la Momie'' (noted ''Momie''). Right: an aerial view of the ''Pré de Madame Carle'' alluvial plain. The downstream landslide is visible in the bottom-right corner of the picture (south-east).

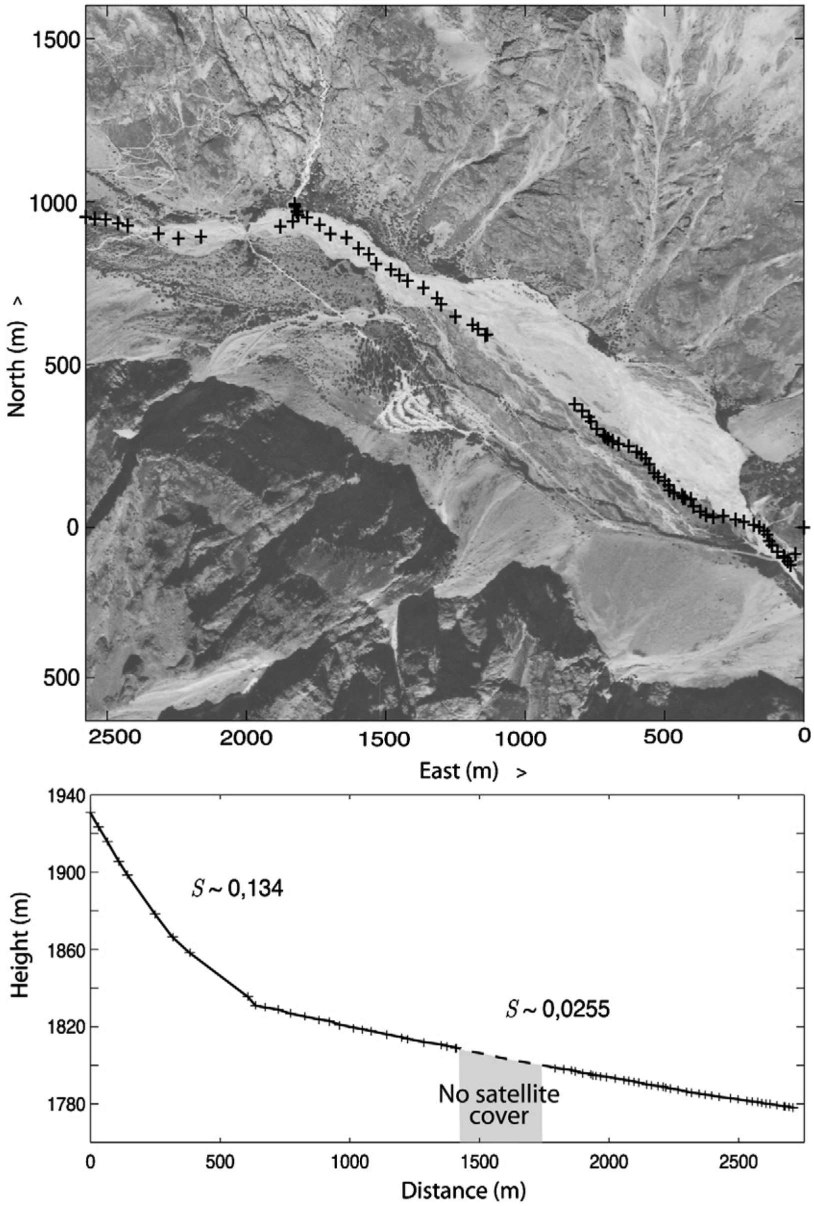

Figure 2 GPS profile of the surface of the stream from the upstream ''torrent du Glacier Noir'' to the outlet of the braiding valley.

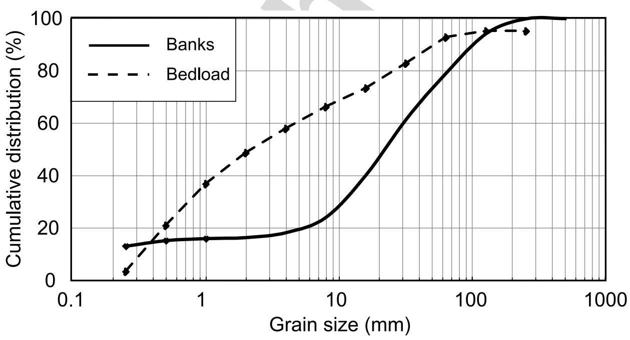

Figure 3 Grains size distribution 1- of the alluvial plain computed from 810 samples collected along the banks and 2- of the bedload material collected over the eight days.

(cobbles, pebbles and granules) with sand and glacial powder (Fig. 3). We used the Wolman (1954) method with a sampling interval of 1 m to determine the grain size distribution of the plain. The samples were collected along the banks, assuming that the highly dynamic geometry of the braided system reduces particle sorting. Thus, we expect few differences in the grain size distribution of the sediment constituting the bed and the distribution of the sediment constituting the banks. We determined that the D_{90} of the material collected along the banks is nearly 9 cm. The torrent, mainly fed by the melting of the glaciers, flows from roughly May to October. The bankfull discharge of the single channel at the outlet of the plain is reached, few hours a day, from around July the 1st to August the 15th. During this period, the weather is generally dry and sunny, with few rainfalls of small magnitude. Climatic data (Météo-France)

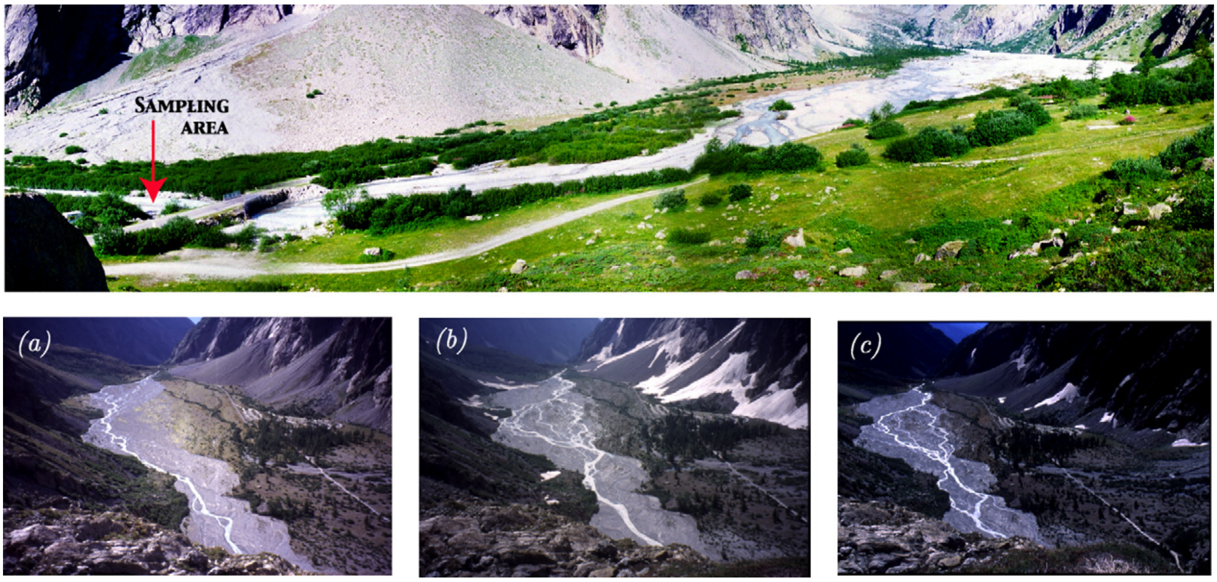

Figure 4 Top: panoramic view of the torrent from the confluence (right) to the measurement point (left) on the 16/07/02. Bottom: three snapshots of the braided network taken at different times: 08/13/99 (a), 07/03/01 (b) and 08/24/01 (c), illustrating the dynamics of the network.

collected in Vallouise (few kilometers downstream), are reported in Appendix. Because of its high dynamics, the river morphology in the plain can completely change over short periods of time (Fig. 4, bottom).

Measurement method

Our measurement site is located at the south of the plain, where the channels merge to form a single straight channel (Fig. 4, top). In this way, we were able to measure the total discharge at the outlet of the braiding system; this has been done three times a day from July the 15th 2002 to July the 22nd 2002. During these 8 days, the weather was sunny, excepted on July the 21nd, where a clouds cover took place above the catchments, reducing the ice melting. The monthly values of temperature, sunshine duration and precipitation for 2002 are reported in Appendix. The flow velocity has been measured with a propeller current meter OTT by establishing sections across the river with a sampling interval of 1 m, that is to say, 10 subsections per section. This number is chosen in order to provide enough subsection to prevent the subsection-discharge being greater than 20% of the total discharge. This reduced number of subsections is imposed by the necessity of performing the gaging over an hour and half while the discharge increased at the rate of 0.3 m^3/s/h. During the sampling period, the channel width varied from 9.2 m the 15th to 11 m the 22nd (Fig. 5c). At subsections, vertical velocity sampling was performed every 10 cm from the bed to the surface, and at 5 cm from the bottom. The sampling time for velocity measurement ranged from 30 s to 1 min. The discharge through the river varied between 2.2 and 4.6 m^3/s.

The suspended load and the dissolved load were collected with a USDH48 integrating water sampler (Guy and Norman, 1970). We generally collected 750 ml of water in the middle of the channel to do our estimations. Suspended load were measured three times a day, and the mid-day sample was used to estimate the dissolved load. The suspended load concentration varies from 0.052 to 7 g/l with a mean of 0.74 g/l and a median of 0.365 g/l. The dissolved load measurements concerns chlorates, nitrates, sulfates and carbonates, sodium, magnesium, potassium and calcium concentrations.[1] Total dissolved load concentration ranges from 35×10^{-3} to 16×10^{-3} g/l.

A Helley–Smith sampler with a 15 cm × 15 cm entrance and 0.25 mm mesh sample bag was used to estimate bed load transport rates (Helley and Smith, 1971). As indicated before, measurements were done at the same place as the velocity measurement and, as much as possible, at the same time (bed load was collected just after the velocity measurement). Sampling duration was 1 min. Bed load catches range from 0 to 4.3 kg with a mean of 0.74 kg and a median of 0.58 kg. The grain size distribution of bed load samples collected over the week is displayed in Fig. 3. The medium diameter is 2 mm but 20% of the bedload is coarser than 25 mm, the medium diameter (D_{50}) of the bed material. This reveals the motion of particles coarse enough to induce changes in the bed configuration.

Results

Mass balance

Fig. 5a presents the total discharge at the outlet of the braiding plain from July the 15th to July the 22nd.[2] The discharge is clearly correlated with the melting of the glaciers. It increases during the day to reach a maximum value at sundown. On the same figure are reported the bed load trans-

[1] Silica measurement will be performed after submission although it will most probably be negligible.

[2] All bed load, velocity measurements and width measurements are available online at http://morpho1.ipgp.jussieu.fr/myGREEC/Publis/Fluvial-_gemorphology/ecrins_data_02.pdf.

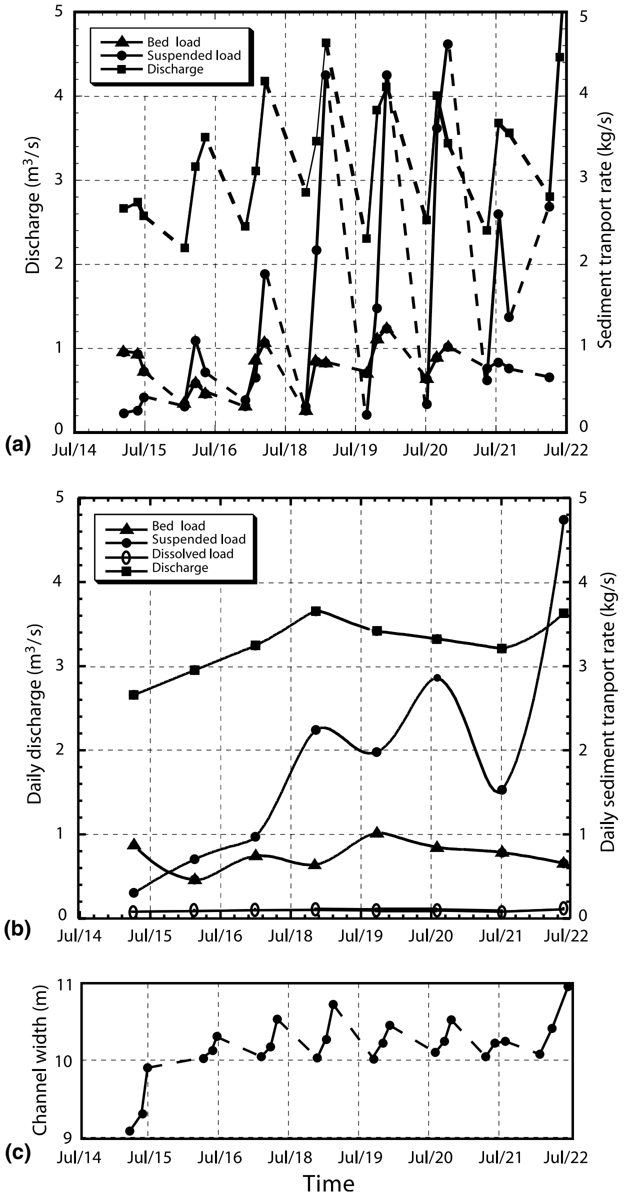

Figure 5 Discharge (a), daily average discharge (b), bed load, suspended load and dissolved load transport rate (b) and channel width (c) at the outlet of the plain from July the 15th to July the 22nd.

port rate and the suspended load transport rate. Each estimation of bed load transport rate corresponds to the average of individual bed load catches over the cross section, multiplied by the wetted perimeter.

In order to establish the mass balance, we report these measurements averaged over each day in Fig. 5b. On July the 15th, the discharge is at its lowest rate and the bed load transport reaches 60% of the total transport rate. Then, the discharge slowly increases (July the 16th and 17th); the bed load transport rate and the suspended load transport rate are of the same order. From July the 18th, the suspended load transport rate increases with discharge to reach a peak of 5 kg/s, becoming thus the main transport mechanism. Meanwhile, the bed load transport rate remains lower than or equal to 1 kg/s. The contribution of the dissolved load, reported in Fig. 5b, remains rather low, reaching a maximum value of 0.12 kg/s. Plotting these data in terms of percentage of the total sediment discharge, the strong variation of the ratio bed load/suspended load becomes obvious (Fig. 6). Bed load transport, which constitutes nearly 60% of the transport at the beginning of the week, eventually falls to less than 15%, while suspended load transport becomes largely dominant. This distribution of transport modes shows clearly that physical erosion is almost the only weathering mechanism at the output of the glaciers.

In a comparable work, Métivier et al. (2004) studied sediment transport in the Ürümqi He, a proglacial stream in the Chinese Tianshan, during the peak flow season (also in July). They found a mass balance as follows: bed load 46% (64% of the solid load), suspended load 25% and dissolved load 29%. The discharge is comparable (1.7 m^3/s) to the case of the torrent de Saint Pierre, although remaining slightly lower. This can be explained by the higher elevation, the drier climate and the smaller size of the Tianshan glaciers. Nevertheless, these results show that both the lithology of the surrounding rocks, and the granulometry of the glacio-fluvial fill feeding the stream, are likely to control of mass balance.

Velocity and sediments transport

Turbulence of the stream

The highly turbulent stream (Reynolds number, $Re = 3.5 \times 10^5$), induces high fluctuations of the velocities we measured despite an integration up to 60 s. Fig. 7 shows the histogram of the Froude number. The large standard deviation of Fr numbers shows the high variability of the flow regime. But it also clearly shows that the distribution presents a well-defined mean value around 1. This suggests that a uniform flow formula may arise from the values measured if the latter is averaged over the fluctuations due to turbulence. The simplest form of this relation is the Chézy formula (Yalin, 1992)

$$\overline{U} = c_f\sqrt{gHs}, \qquad (1)$$

where H is the water height, s is the slope of the water surface and c_f is the friction coefficient. Fig. 8 shows the Chézy plot applied to our data. Our data have been averaged using a binning method over intervals including 15 values. The river regime follows, on average, the Chézy law with an effective friction coefficient of $\simeq$4.4. However, the important

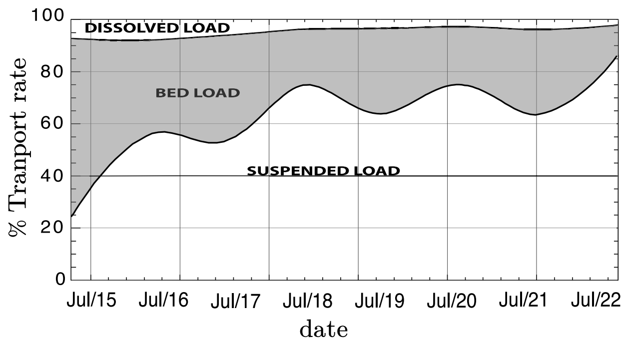

Figure 6 Contribution of bed load, suspended load and dissolved load to the total transport rate at the outlet of the plain from July the 15th to July the 22nd.

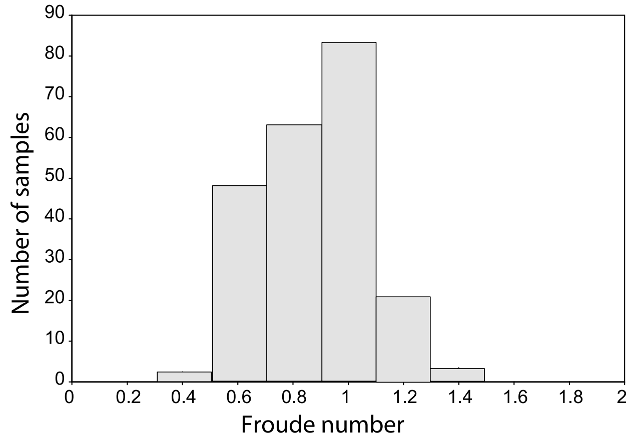

Figure 7 Histogram of Froude numbers *Fr* computed from 225 measurements.

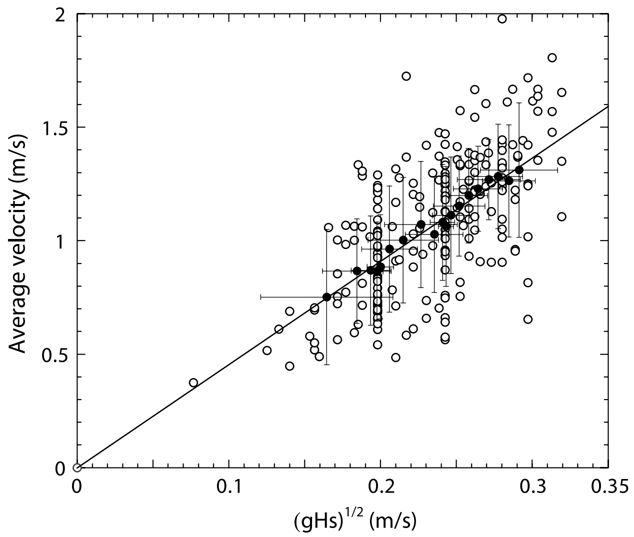

Figure 8 Average velocity plotted against $\sqrt{gHS}$. The linear regression shows that velocity follows the classical Chézy formula.

dispersion of the data illustrates the high turbulence of the flow and the changes in flow regime with time. It is important to notice that in spite of the high variation of the velocity, its average value remains clearly correlated to $(gHs)^{1/2}$.

Relevant velocity

Finding relations between sediment transport and water flow leads to consider which parameter, computed from the velocity data, is relevant for the transport description. One of these parameters is the shear velocity (Bagnold, 1973; Raudkivi, 1990) given by

$$u^* = \sqrt{\frac{\tau_0}{\rho}}, \tag{2}$$

where τ_0 is the shear stress applied to the bed and ρ is the density of water. There are different way to evaluate the shear velocity if one assumes that, at the first order, velocity profiles of the river follow the logarithmic law of the wall (Fig. 9). If so, the shear velocity can be evaluated using the relation

$$\frac{u}{u_p^*} = \frac{1}{\kappa} \ln\left(\frac{z}{z_0}\right), \tag{3}$$

where u is the velocity at a given height z, u_p^* is the shear velocity, $\kappa = 0.4$ is the von Karman constant and z_0 the depth where velocity vanishes. In Eq. (3), u_p^* is directly computed from data by adjustment of this equation without the need of any prediction on the value of z_0. Using the same hypothesis on the velocity profile, one can deduce the shear velocity (now noted u_h^*) from the depth averaged velocity relation, namely

$$\frac{\overline{U}}{u_h^*} = \frac{1}{k} \ln\left(\frac{h}{ez_0}\right), \tag{4}$$

where $\overline{U}$ is the average velocity and h is total height. The value of z_0 can be estimated from the relation

$$z_0 = aD_{84}, \tag{5}$$

provided by many studies on gravel bed rivers (Whiting and Dietrich, 1991; Wiberg and Dungan Smith, 1991; Wilcock, 1996). Wiberg and Smith (1991) showed that Eq. (4) can be used for a range of cases where the ratio D_{84}/h (where h is the height of water) remains lower than unity. In the case of the torrent de St Pierre, this ratio ranges from 0.1 to 0.7. If the velocity profile follows a logarithmic form, both velocities must be strongly correlated. Fig. 10 shows correlation between u_h^* and u_p^* computed from the complete set of velocity profile measured on the torrent. According to numerous works (Wiberg and Smith, 1991; Wilcock, 1996), we use a value of 0.09 for a and $D_{84} = 9$ cm (Fig. 3), that gives $z_0 = 0.008$ m. The absence

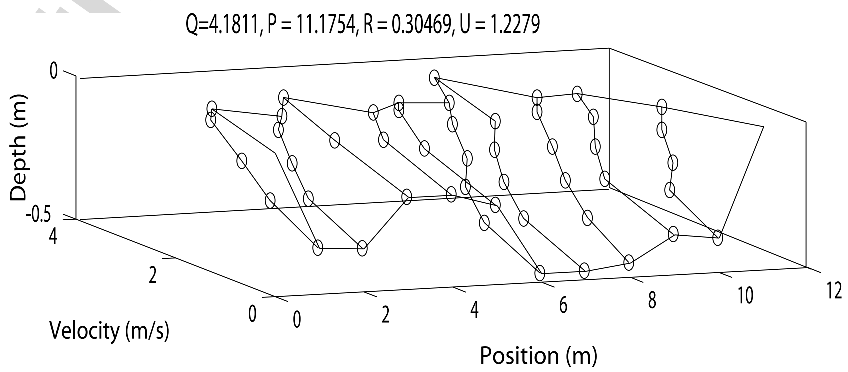

Figure 9 An example of velocity profiles in a cross section.

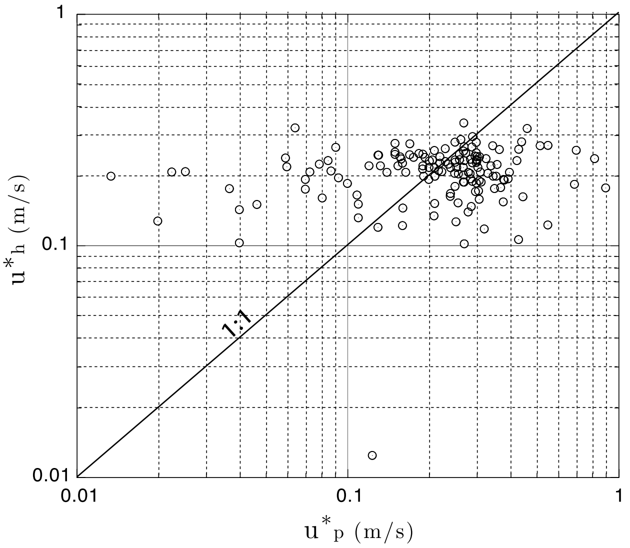

Figure 10 Different estimates of the shear velocity. U_h^* is computed from the average velocity profiles (Eq. (4)), U_p^* is computed by fitting the law of the wall (Eq. (3)).

of any correlation cannot be explained by the duration of the sampling time. Indeed, we have shown that this sampling duration (1 min) is sufficient to draw uniform flow relationships from the data and that the average velocity measured is relevant. This sampling time, very long with regards to high frequency fluctuations of the velocity field, also corresponds to the sampling time of bed load, and hence, to the velocity of the flow that shears the bed during sampling.

Calculation of the shear velocity by means of logarithmic law adjustments is therefore inadequate in very shallow and turbulent mountain streams as the torrent de St Pierre. Because of its lack of meaning, this technique just adds uncertainties to a potential transport law.

The only valuable relation is obtained by plotting the transport rate as a function of the average velocity $\overline{U}$, computed $\overline{U}$ using a bining method. Looking at the correlation in Fig. 11, a power-law relationship can be fitted with an exponent varying from 2 to 4, depending on the size of the bining interval. The highest value of the correlation coefficient is 0.72, obtained for an interval of 10 points, which corresponds to a power law exponent of $\simeq 3$.

However, a problem appears when the velocity vanishes while bed load transport does not. This ''residual transport'' raises the problem of averaging for small values of the velocity. In case of low velocities, the variations become high compared to average values, and can generate temporary conditions of grain movement. Another possible explanation for this residual bed load transport at zero velocity is intrinsic errors induced by the use of Helley—Smith sampler. Indeed, a part of bed load may enter the sampler under the effect of the shock caused by the sampler on the bed. The relative importance of this error may become high when bed load rate is low.

The relation obtained is close to the Bagnold formulation for the bed load transport over the total flow width W (Bagnold, 1973)

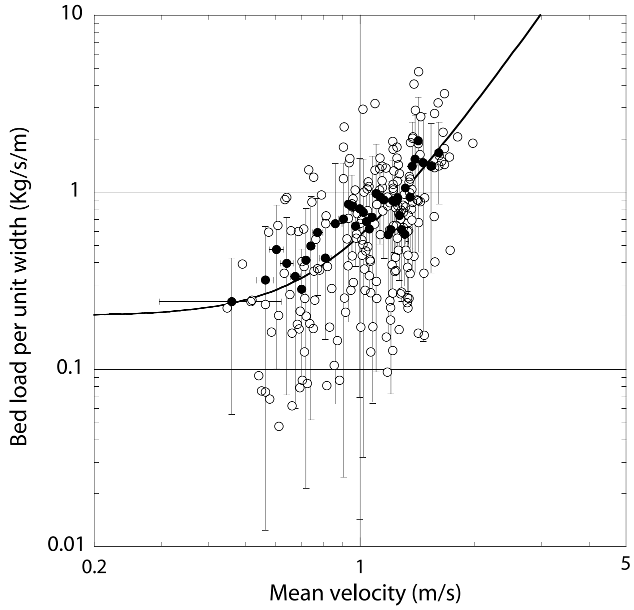

Figure 11 Bed load transport rate per unit width as a function of the average velocity $\overline{U}$. The best correlation is obtained for a power-law relation with an exponent to 3 ± 1.

$$\frac{Q_s}{W} \propto (\tau_* - \tau_{*c})^{3/2} \propto u_*^3 \qquad (6)$$

(where τ_* and τ_{*c} are, respectively, the shear stress and the critical shear stress required for the induction of movement) with the shear velocity replaced by the average velocity

$$\frac{Q_s}{W} \propto \overline{U}^3. \qquad (7)$$

Using the expression of the discharge per unit width

$$\frac{Q}{W} = \overline{U}H, \qquad (8)$$

where H is the water height, Eq. (3) allows to express the discharge as follows:

$$\frac{Q}{W} = c\sqrt{gHs}H \propto s^{1/2}H^{3/2}. \qquad (9)$$

Eq. (6) can thus be reformulated as

$$Q_s \propto c^3 g^{3/2} s^{3/2} H^{3/2} W \propto sQ \propto \Omega, \qquad (10)$$

where Ω is the total stream power of the river. This result is in good agreement with bed load description provided by experimental results on micro-scaled braided rivers (Meunier and Métivier, 2000; Métivier and Meunier, 2003; Meunier and Métivier, 2006). In fact, the use of the average velocity allows to filter out the strong turbulence effects and to link experiment results and field measurements. However, the information contained in dispersion may be useful too. As turbulence effects are related to bed grain size and bed load grain size, all of our bed load samples have been returned to the laboratory. A study is currently in progress to quantify the bed load transport per grain size and relate it to the velocity variations. A closer relation with laboratory studies cited above could be established by carrying a differential fluxes measurement. Indeed, Métivier and Meunier (2003) and Meunier and Métivier (2006) suggest that the bedload fluxe measured at a given cross section is a function of its value at the inlet of the braiding plain and the distance between the cross section and the inlet. Our future field works on the torrent de St Pierre will include measurement at different locations along the alluvial plain to enforce that idea.

Relation between bed load and suspended load

In this section, we focus on a promising correlation between suspended load and bed load fluxes. Indeed, it is possible to correlate the two fluxes at a given cross section. The suspended load transport rate is plotted against the bed load transport rate in Fig. 12. Using a binning method, the relation on average, seems to follow a power-law relationship with an exponent $\simeq 4$. Here again, the dispersion due to the temporal and spatial variability, is important and shows the necessity of using a height number of measurements. The existence of a correlation between both fluxes at a given cross section implies that a significant portion of the suspended load originates from the bed, and not directly from the glaciers. It also leads to consider that the system is close to the equilibrium, that is to say that the amount of fine particles released by moving coarser grains is constant along the stream.

In the case of low external flux (coming from the slopes for example), the suspended material can be considered as coming from either the river bed or directly from the glacier. This assumption can be made for the torrent de St Pierre. Indeed, the short distance between the sampling area and the glacier front (corresponding to a small drainage area), and the low value of precipitations during the week allows us to assume external fluxes to be negligible. Thus, sediments in suspension all originate from the glacier. No continuous survey of suspended matter at the glacier limit was done but two single water samples were taken in order to constraint the order of magnitude on the concentration originating from the glaciers. The concentration of the suspended load is close to 7 g/l for the torrent du Glacier Noir and 1 g/l for the torrent du Glacier Blanc. When compared with the average concentration of 0.72 g/l and the median concentration of 0.365 g/l measured at the survey site these high values clearly suggest that sedimentation takes place on the alluvial plain. If not, there would be no reason for this correlation since the suspended load would never be in contact with the bed material. One can guess that the variability of measurements comes from the portion of suspended load that directly originates from the glaciers.

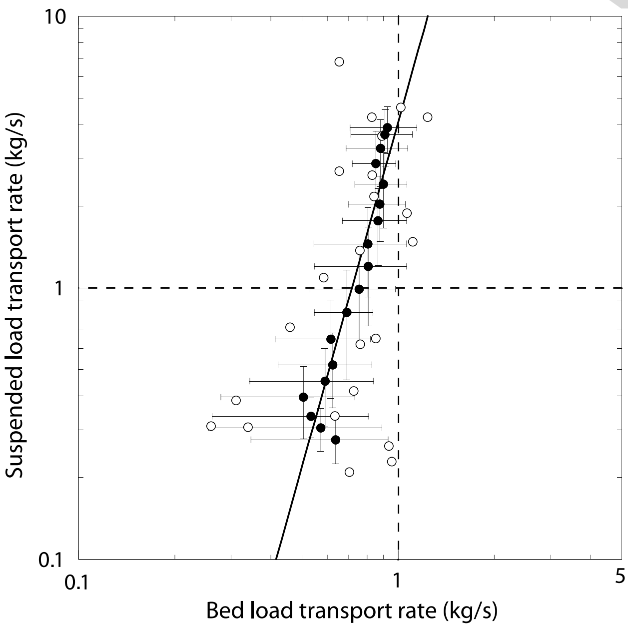

Figure 12 Suspended load transport rate plotted against the bed load transport rate. The correlation shows a power-law relationship with an exponent of 4.

Concerning the suspended flux coming from the river bed, the process can be qualitatively described as follows: the motion of coarser grains implies a release of fine particles trapped in the bed. If the suspended load moves at stream velocity, the suspended load flux measured at a given cross section should depend on the amount of coarser grains in movement integrated over an upstream segment of the river. The length of this segment, over which the suspended flux is integrated, yet remains unknown.

This kind of ''flushing process'' has also been observed in another proglacial stream, the Ürümqi He, in the Chinese Tian Shan. A correlation between average daily bed load and suspended load was established along the river from the glacier to the range front over 60 km (Métivier et al., 2004). Both fluxes are also related by a power law, but in the case of the Ürümqi He, bed load flux is prevalent and the exponent is lower than one (Métivier et al., 2004). Further measurements of suspended flux along the river from the glacier front to a downstream point will be to confirm the suspended flux/bed motion correlation, and could provide a quantitative evaluation of the partial flux coming from glacier themselves. They also would enable the study of the liability of such relation in case of larger runoff. Indeed, the greater the discharge, the more rare the interaction between suspended particles and the bed. Thus, one can expect a degradation of the correlation (or a change in the value of the exponent) if the bed load does not increase enough to release a significant amount of fine particles to maintain the relationship.

Conclusion

Measurements made in the torrent St Pierre in high flow season show that the river load is mainly of solid matter transport. The proglacial nature of this river results in a large amount of material ranging from glacial powder to boulder size particles that results from abrasion of rocks by the overlying glaciers. Solid load is mainly composed of sediment in suspension. However, we measured that bed load fluxes are far from being negligible and can be prevalent at low discharge. By sampling velocity and bed load over short durations, we obtained instantaneous quantities that do not demonstrate a relation between the bed load and the shear velocity. The only relevant parameter for bed load transport description is shown to be the average velocity. The power-law relationship obtained, when expressed in term of discharge, is in good agreement with laboratory results (Meunier and Métivier, 2000; Métivier and Meunier, 2003). We shown that in the torrent de Saint Pierre, during the snowmelt runoff, suspended load transport is correlated to bed load transport. The existence of such a relation between these fluxes suggests that the majority of particles in suspension come from the bed. Hypothesis required to the prevalence of such a mechanism can be done in the case of this proglacial river because the external fluxes are not substantial over the drainage area considered. This kind of relation is very interesting because of the possible evaluation of the bed load rate without complicated and uncertain field measurements. However, further measurements, focused on the suspended load evolution along the plain, are necessary to highlight this process.

Acknowledgements

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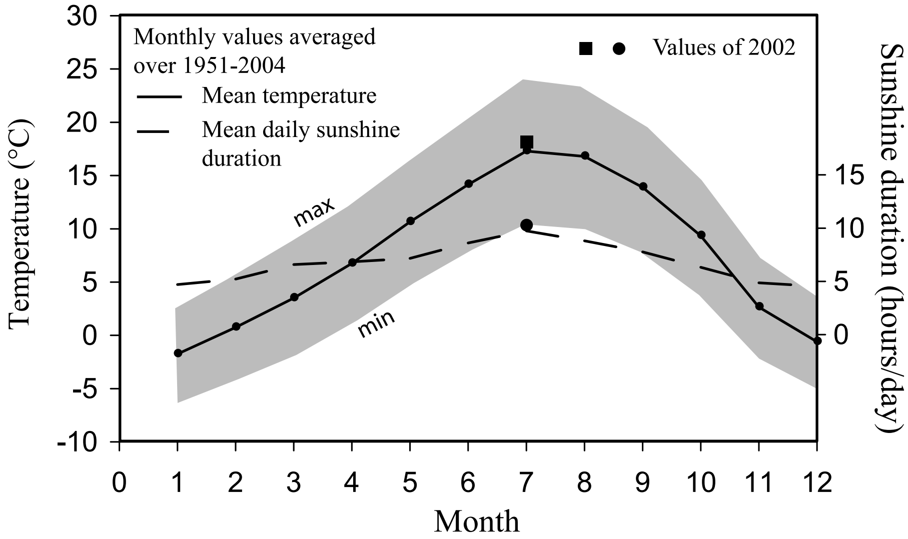
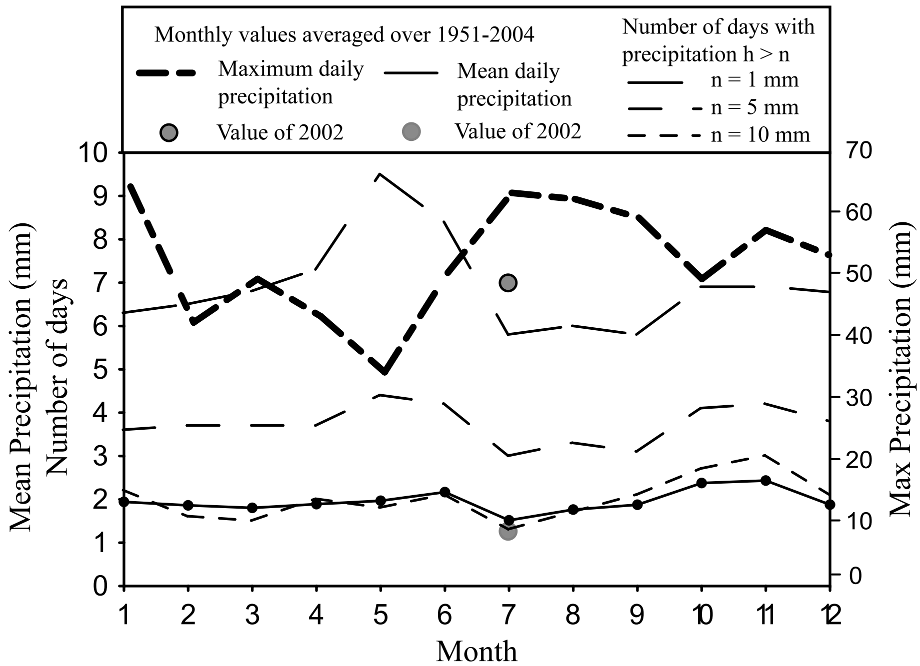
Appendix





Meteorological measurements in Vallouise, located few kilometers downstream (source: Météo-France).

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