Short communication: Field data imply that the sorting (D96/D50 ratios) of grains on fluvial gravel bars influences the probability of sediment entrainment

Running title: transport probability of coarse-grained material

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Abstract
Conceptual models suggest that the mobility of grains on coarse-grained gravel bars is mainly controlled by sediment supply. Here we present field observations from streams in the Swiss Alps and the Peruvian Andes to document that for a given water runoff, the probability of material transport also depends on the sorting of the bed material. We calculate shear stresses that are expected for a mean annual water discharge, and compare these estimates with grain-specific thresholds. We find a positive correlation between the predicted probability of material transport and the sorting of the bed material, expressed by the D96/D50 ratio. These results suggest that besides sediment supply, the bedload sorting exerts a measurable control on the mobility of clasts in coarse-grained streams.

1 Introduction
It is generally accepted that sediment supply is one of the most important parameters, which not only controls the mobility of the sediment in coarse-grained streams but also the channel form (Dade and Friend, 1998; Church, 2006). In particular, flume experiments (Dietrich et al., 1989) and numerical models (Wickert et al., 2013) have shown that a large sediment supply is commonly found in braided rivers where the material mobility is high, while a low sediment mobility is rather encountered in single-threat channels where the sediment supply is expected to be low. However, much less research has been conducted to investigate whether the granulometric composition, and particularly the sorting of the bed material, also exerts a measurable control on the mobility of coarse-grained material in streams. Here, we focus on this aspect and explore whether there is a link between the sorting of gravel bars, here expressed by the D96/D50 ratios of the material, and the transport probability of individual clasts on these bars. We focus on gravelly streams in the Swiss Alps where artificial banks keep the flow in fixed, single-threat channels over several kilometres, and in the Peruvian Andes where streams are braided. We selected gravel bars close to water gauging stations, determined the grain size distribution of these bars and calculated the probability of sediment transport for a selected water runoff, which in our case corresponds to the mean annual water discharge Q_{mean} for comparison purposes. We explored whether these flows are strong enough to shift the D96 grain size, which is considered to build the sedimentary framework of gravel bars as recent flume experiments have shown (MacKenzi and Eaton, 2017;
2 Methods and datasets

2.1 Entrainment of bedload material

Sediment mobilization is considered to occur when flow strength $\tau$ exceeds a grain size specific threshold $\tau_c$ (e.g., Paola et al., 1992):

$$\tau > \tau_c$$

We estimated the probabilities of $\tau > \tau_c$ for a given water discharge using a Monte Carlo modeling framework (see section 2.2). We conducted 10’000 simulations, and the results are reported as probability (or percentage) where $\tau > \tau_c$ during these iterations.

Threshold shear stress $\tau_c$ for the dislocation of grains with size $D_i$ can be obtained using Shields (1936) criteria $\phi$ for the entrainment of sediment particles:

$$\tau_c = \phi \left( \rho_s - \rho \right) g D_i$$

where $g$ denotes the gravitational acceleration, and $\rho_s$ (2700 kg/m³) and $\rho$ the sediment and water densities, respectively. Among the various grain sizes, the 84th percentile $D_{84}$ has been considered to best characterize the sedimentary framework of a gravel bar (Howard, 1980; Hey and Thorne, 1986; Grant et al., 1990). Accordingly, flows that dislocate the $D_{84}$ grain size are strong enough to alter the gravel bar architecture (Grant et al., 1990). We acknowledge that many authors preferentially selected the $D_{50}$ grain size as a threshold to quantify the minimum flow strengths $\tau_c$ to entrain the bed material (e.g., Paola and Mohrig, 1996; Pfeiffer and Finnegan, 2018). The selection of the $D_{50}$ thus results in relatively low thresholds and in a greater entrainment probability. However, recent analogue experiments have shown that the coarse-grained trail of a material composition such as e.g., the $D_{50}$ better characterizes the threshold conditions for the incipient motion of material on gravel bars than the $D_{50}$ (MacKenzie and Eaton, 2017). We therefore followed the recommendations by MacKenzie et al. (2018) and selected the $D_{84}$ grain size to quantify the...
Bed shear stress $\tau$ is computed through (e.g., Tucker and Slingerland, 1997):

$$\tau = \rho g RS$$  \hspace{1cm} (3).

Here, $S$ denotes the energy gradient, and $R$ is the hydraulic radius, which is approximated through water depth $d$ where channel widths $W > 20 \times d$ (Tucker and Slingerland, 1997), which is the case here. The combination of expressions for: (i) the continuity of mass including flow velocity $V$, channel width $W$ and water discharge $Q$:

$$Q = VWd$$  \hspace{1cm} (4);

(ii) the relationship between flow velocity and channel bed roughness $n$ (Manning, 1891):

$$V = \frac{1}{n} d^{2/3} S^{1/2}$$  \hspace{1cm} (5);

and (iii) an equation for the Manning’s roughness number $n$ (Jarrett, 1984):

$$n = 0.32 S^{0.38} d^{-1/6}$$  \hspace{1cm} (6);

yields a relationship where bed shear stress $\tau$ depends on gradient, water flux and channel width (Litty et al., 2017):

$$\tau = 0.54 \rho g \left( \frac{Q}{W} \right)^{0.55} S^{0.935}$$  \hspace{1cm} (7).

This equation is similar to the expression by Hancock and Anderson (2002), Norton et al. (2016) and Wickert and Schildgen (2019) with minor differences regarding the exponent on the channel gradient $S$ and on the ratio $Q/W$. These mainly base on the different ways of how bed roughness is considered. Note that this equation does not consider a roughness length scale (both vertical and horizontal) because we have no constraints on this variable.

We explored whether equation (2) could be solved using the Darcy-Weisbach friction factor $f$ instead of Manning’s $n$. According to Ferguson (2007), the friction factor $f$ varies considerably between shallow- and deep-water flows and depends on grain size $D_s$ relative to water depth $d$, and thus on the relative roughness. Ferguson (2007) developed a solution referred to as the Variable Power Equation (VPE), which accounts for the dependency of $f$ on the relative importance of roughness-layer versus skin friction effects and thus on the $D_s/d$ ratios. Calculations where the VPE was employed indeed revealed that roughness-layer effects have an impact on flow regimes where $D_s/d > 0.2$ (Schlunegger and Garefalakis, 2018), which is likely to be the case in our streams. However, similar to Litty et al. (2016), we are faced with the problem that we have not sufficient constraints to analytically solve equation (2) with the VPE. We therefore selected Mannings’s $n$ instead, which allowed us to solve equation (2) analytically. But we acknowledge that this might introduce a bias.

2.2 Monte Carlo simulations
Predictions of sediment transport probability are calculated using Monte Carlo simulations performed within a MATLAB computing environment. All variables that are considered for the calculations of both sheer and critical sheer stresses (equations 7 and 2, respectively) are randomly selected within their possible ranges of variation (see next sections and Table 1). Except for the Shields $\phi$ variable that we consider to follow a uniform distribution between 0.03 and 0.06 (see section 2.3.1 for justification), we infer that all other variables follow a normal distribution, defined by its mean and one standard deviation. To ensure that no negative values introduce a bias to these iterations, only strictly positive values for channel widths and gradients are considered. In the case of water discharge, both null and positive values are kept for further calculations. Values excluded from the calculations, i.e. returning negative water discharge or null or negative channel width / slope gradient, yield “NaN” in the resulting vector. For each of the 10’000 iterations $\tau$ and $\tau_c$ are compared, which yields either “1” ($\tau > \tau_c$) or “0” ($\tau \leq \tau_c$). The sediment transport probability is then calculated as the sum of ones divided by the number of draws, from which the number of “NaN” values was subtracted before. Note that <2500 “NaN” were obtained for Rio Chico (PRC-ME17), which we mainly explain by the c. 150% relative standard deviation of the mean annual water discharge estimated for that river.

2.3 Parameters, datasets, uncertainties and sensitivity analysis

2.3.1 Shields variable $\phi$

Assignments of values to $\phi$ vary and diverge between flume experiments (e.g., Carling et al., 1992; Ferguson, 2012; Powell et al., 2016) and field observations (Mueller et al., 2005; Lamb et al., 2008). Here, we considered that at the incipient motion of the $D_{84}$, the Shields variable $\phi$ is equally distributed between 0.03 and 0.06 (Dade and Friend, 1998) during the 10’000 iterations. However, based on a compilation of $\phi$-values that were derived from field investigations and flume experiments, Lamb et al. (2008) revealed that $\phi$ is likely to depend on the energy gradient itself, where

$$\theta = 0.15S^{0.25}$$

We refrain from using a slope dependency of $\phi$ at this stage for three major reasons: First, the consideration of equally distributed $\phi$-values, in the range between 0.03 and 0.06, includes the Shields values of most field investigations and flume experiments where channel gradients were between 0.001 and 0.02 (spread of energy gradients of our streams, Table 1), as Lamb et al. (2008) and Bunte et al. (2013) have shown in their compilations. Second, $\phi$-values show a large scatter in their slope-dependencies (Lamb et al., 2008). Accordingly, the consideration of equally distributed $\phi$-values within a given range better complies with the large variability in $\phi$-values that are commonly encountered in experiments and field surveys (Lamb et al., 2008). Third, the selected range considers most of the complexities that are related to the hiding of small clasts and the protrusion of large constituents (Buffington et al, 1992; Buffington and Montgomery, 1997;
Kirchner et al., 1990; Johnston et al., 1998), which, in turn, results in a large scatter of $\phi$-values. In this context, Turowski et al. (2011) reported a larger variation in the threshold conditions for the mobilization of clasts than employed here. However, their streams have energy gradients between 0.06 and 0.1, with the consequence that some of the material is entrained during torrential floods where entrainment mechanisms are different. Finally, the selected $\phi$-range also includes the hydrological conditions of channel forming floods where thresholds for the evacuation of sediment may be up to 1.2 times larger than for the incipient motion of individual clasts (Parker, 1978; Philips and Jerolmack, 2016; 2019; Pfeiffer et al., 2017). For instance, a 1.2-times larger threshold will increase the commonly employed $\phi$ value of 0.047 (Meyer-Peter and Müller, 1948), or alternatively 0.0495 (Wong and Parker, 2006), to the range between 0.036 and 0.0594, which is considered in the brackets of 0.03 and 0.06 that we employed in this paper. In summary, we consider that the selection of equally distributed $\phi$-values between 0.03 and 0.06 does the best job to account for the large variability in $\phi$-values that are commonly encountered in experiments and field surveys where energy gradients were between 0.001 and 0.02 (Lamb et al., 2008).

2.3.2 Grain size data

We collected grain size data from streams where water discharge has been monitored during the past decades. These are the Kander, Lütschine, Rhein, Sarine, Simme, Sitter and Thur Rivers in the Swiss Alps (Fig. 1a). The target gravel bars are situated close to a water gauging station. At these sites, 5 to 6 digital photographs were taken with a Canon EOS PR. The photos covered the entire lengths of these bars. A meter stick was placed on the ground and photographed together with the grains. Grain sizes were then measured with the Wolman (1954) method using the free software package ImageJ 1.52n (https://imagej.nih.gov). Following Wolman (1954), we used intersecting points of a grid to randomly select the grains to measure. A digital grid of 20x20 cm was calibrated with the meter stick on each photo. The size of the grid was selected so that the spacing between intersecting points was larger than the $b$-axis of most of the largest clasts (Table 1, Supplement S1). The grid was then placed on the photograph with its origin at the lower left corner of the photo. The intermediate or $b$-axis of approximately 250 – 300 grains (c. 50 grains per photo; Supplement S1) underneath a grid point was measured for each gravel bar. In this context, we inferred that the shortest ($c$-axis) was vertically oriented, and that the photos displayed the $a$- and $b$-axis only. In cases where more than half of the grain was buried, the neighboring grain was measured instead. In the few cases where the same grain lay beneath several grid points, then the grain was only measured once. Only grains larger than a few millimeters (>4-5 mm, depending on the quality of the photos) could be measured. While the limitation to precisely measure the finest-grained particles potentially biases the determination of the $D_{50}$, it will not influence the measurements of the $D_{84}$ and $D_{90}$ grain sizes, as the comparison between sieving and measuring of grains with the Wolman (1954) method has disclosed (Watkins et al., 2020). In addition, as will be shown below, the consideration of the
instead of the $D_{96}/D_{50}$ ratios yields a similar positive relationship to the mobility of grains. We complemented the grain size data sets with published information on the $D_{50}$, $D_{84}$ and $D_{96}$ grain size (Litty and Schlunegger, 2017; Litty et al., 2017) for further streams in Switzerland and Peru (Figs. 1a and 1b; Table 1). For a few streams in Switzerland, Hauser (2018) presented $D_{84}$ grain size data from the same gravel bars as Litty and Schlunegger (2017), but the photo was taken one year later and possibly from a different site. For these 5 locations, we took the arithmetic mean of both surveys (Table 1, data marked with three asterisks). All authors used the same approach upon collecting grain size data, which justifies the combination of the new with the published datasets.

We finally assigned an uncertainty of 20% to the $D_{84}$ threshold grain size, which considers the variability of the $D_{84}$ within a gravel bar as the analysis of the intra-bar variation of the $D_{84}$ for selected gravel bars in Switzerland shows (Supplement S1). The assignment of a 20% uncertainty to the $D_{84}$ threshold grain size also considers a possible bias that could be related to the grain size measuring technique (e.g., sieving in the field versus grain size measurements using the Wolman method; Watkins et al., 2020). However, it is likely to underestimate the temporal variability in the grain size data, as a repeated measurement on some gravel bars in Switzerland has suggested (Hauser, 2018), but this aspect warrants further research.

2.3.3 Water discharge data

The Federal Office for the Environment (FOEN) of Switzerland has measured the runoff values of Swiss streams over several decades. We employed the mean annual discharge values over 20 years for these streams (Supplement S2) and calculated one standard deviation thereof (see Table 1). For the Peruvian streams, we used the mean annual water discharge values $Q_{mean}$ reported by Litty et al. (2017) and Reber et al. (2017). These authors obtained the mean annual water discharge (Table 1) through a combination of hydrological data reported by the Sistema Nacional de Información de Recursos Hídricos and the TRMM-V6.3B43.2 precipitation database (Huffman et al., 2007). They also considered the intra-annual runoff variability as one standard deviation from $Q_{mean}$ to account for the strong seasonality in runoff for the Peruvian streams, which we employed in this paper. For the Peruvian streams, the assigned uncertainties to $Q_{mean}$ are therefore significantly larger than for the Swiss rivers (Table 1). A re-assessment of the inter-annual variability of water discharge for those streams in Peru where the gauging sites are close to the grain size sampling location (distance of a few kilometers) yields a one standard deviation of c. 50%, which is still much larger than for the Swiss rivers (Appendix S2). We therefore run sensitivity tests where we considered scenarios with different relative values for 1σ standard deviations of $Q_{mean}$.

We additionally ran sensitivity tests to explore how the mobility probability changes if runoff quantiles instead of $Q_{mean}$ are considered (Supplement S3). In particular, we ran a series of Monte Carlo simulations for various runoff quantiles and then calculated the resulting probability of
sediment mobilization for each of these quantiles. We then multiplied the occurrence probability of each runoff quantiles (listed by the Swiss authorities and calculated for the Peruvian streams based on 4 to 98-years equivalent daily records) with the corresponding transport probability and summed the values. This integration provides an alternative and more realistic estimate of transport probability (Supplement S3).

2.3.4 Channel width data
For the Swiss streams, channel widths and gradients (Table 1, Supplement S4) were measured on orthophotos and LiDAR DEMs with a 2-m resolution provided by Swisstopo. From this database, gradients were measured over a reach of c. 250 to 500 m. All selected Swiss rivers are single-thread streams following the classification scheme of Eaton et al. (2010), and flows are constrained by artificial banks where channel widths are constant over several kilometers. For these streams, we therefore measured the cross-sectional widths between the channel banks, similar to Litty and Schlunegger (2017).

We complemented this information with channel width (wetted perimeter) and energy gradient data for 21 Peruvian streams that were collected by Litty et al. (2017) in the field and on orthophotos taken between March-June. This period also corresponds to the season when the digital photos for the grain size analysis were made (Mai 2015). We acknowledge that widths of active channels in Peru vary greatly on an annual basis because of the strong seasonality of runoff (see above and large intra-annual variability of runoff in Table 1). We therefore considered scenarios where channel widths are twice as large as those reported on Table 1 (see Supplement S5).

The uncertainties on slope and channel width largely depend on the resolution of the digital elevation models underlying the orthophotos (2-m LiDAR DEM for Switzerland, and 30-m ASTER DEM for Peru). It is not possible to precisely determine the uncertainties on the slope values. Nevertheless, we anticipate that these will be smaller for the Swiss rivers than for the Peruvian streams mainly because of the higher resolution of the DEM. We ran sensitivity models where we explored how the probability of material transport changes in the Swiss rivers for various uncertainties on channel widths, energy gradients and mean annual discharge values (Supplement S4).

3 Results
The grain sizes range from 8 mm to 70 mm for the $D_{50}$, 29 mm to 128 mm for the $D_{84}$, and 52 mm and 263 mm to the $D_{96}$. The smallest and largest $D_{50}$ values were determined for the Maggia and Rhein Rivers in the Swiss Alps, respectively (Table 1). The grain sizes in the Swiss Rivers also reveal the largest spread where the ratio between the $D_{95}$ and $D_{50}$ grain size ranges between 2.2 (Sarine) and 17.7 (Maggia Losone I), while the corresponding ratios in the Peruvian streams are
between 2.1 (PRC-ME9) and 5.8 (PRC-ME17). In the Swiss Alps, the critical shear stresses \( \tau_c \) (median values) for entraining the \( D_{50} \) grain size ranges from c. 20 Pa (Emme River) to c. 90 Pa (Rhein and Simme Rivers). In the Peruvian Andes, the largest critical shear values are <80 Pa (PRC-ME39). The shear stress values related to the mean annual water discharge \( Q_{mean} \) range from c. 15 Pa to 100 Pa in the Alps and from 20 Pa to >400 Pa in the Andes. Considering the strength of a mean annual flow and the \( D_{50} \) grain size as threshold, the probability of sediment transport occurrence in the Peruvian Andes and in the Swiss Alps comprises the full range between 0% and 100%.

Rivers that are not affected by recurrent high magnitude events (e.g., debris flows) and where the grain size distribution is not perturbed by lateral material supply are expected to display a self-similar grain size distribution (Whittaker et al., 2011; D’Arcy et al., 2017; Harries et al., 2018), characterized by a linear relationship between the \( D_{50}/D_{90} \) and \( D_{50}/D_{95} \) ratios. In case of the Maggia River, the largest grains are oversized if the \( D_{50} \) and the grain size distribution of the other streams are considered as reference (Fig. 2). This could reflect a response to the supply of coarse-grained material by a tributary stream where the confluence is <1 km upstream of the Maggia sites. Alternatively, and possibly more likely, it reflects the response to the high magnitude floods in this stream (Brönnimann et al., 2018). In particular, while the ratio between the last and first quantiles is <150 in the Swiss streams on the northern side the Alps, the ratio is 860 in the Maggia River. Interestingly, such ratios are not rare in Peru. However, we anticipate that the Peruvian streams are capable to accommodate such large runoff variabilities through much wider channel belts that are not confined by artificial banks along most of the streams. If we exclude the Maggia dataset, then the probability of sediment transport occurrence scales positively and linearly with the \( D_{90}/D_{50} \) ratios (Fig. 3A). The observed relationship appears stronger for the Swiss rivers (\( R^2 = 0.74 \), p-value = 2E-4) than for the Peruvian stream (\( R^2 = 0.33 \), p-value = 4E-3). These correlations suggest that gravel bars with a poorer sorting of the bed material, here expressed by a high \( D_{90}/D_{50} \) ratio, have a greater probability for the occurrence of sediment transport than those with better-sorted material. If the normalized residuals are plotted against the sorting, then they do not show any specific and significant patterns, and therefore appear independent of the sorting (Fig. 3B). This suggests that the inferred linear relationships between the probability of transport occurrence and the \( D_{90}/D_{50} \) are statistically robust. Although Fig. 3A suggests that the regression for the Swiss rivers (slope= 0.16±0.06; intercept= -0.34±0.31) differs from that of the Peruvian streams (slope= 0.18±0.11; intercept= -0.02±0.46), the regression parameters do not significantly differ when considering them within their 95% confidence intervals.

The use of discharge quantiles yields sediment transport probabilities within a large range between <10% and >50%, and they are higher than reported values from other streams (Torizzo and Pitlick, 2004; Pfeiffer and Finnegan, 2008). Additionally, the resulting probabilities are positively and linearly correlated with the probability of transport estimated with \( Q_{mean} \) (Figure S3 in Supplement), and the correlations are very similar between the Swiss (slope: 0.74±0.02; intercept: 0.05±0.01) and
Peruvian streams (slope: 0.73±0.19; intercept: 0.03±0.14). The mean annual discharge estimates $Q_{\text{mean}}$ are likely biased by infrequent, but large magnitude floods, which could explain the 25% larger transport probabilities if $Q_{\text{mean}}$ is used as reference runoff.

The assignments of different uncertainties on reach slopes, channel widths and discharge has no major influence on the inferred relationships between transport probability and sorting (Supplement S4, S5). For the Peruvian streams, however, assignments of twofold larger values to channel widths will decrease the probability of transport for a given sorting by c. 10-15%. The inferred linear relationship between both variables, however, will remain (Supplement S5).

4 Discussion and Conclusions

The sediment transport calculation is based on the inference that water discharge is strong enough to entrain the frame building grain size $D_{50}$. Therefore, the relationships between the mobilization probability and the $D_{90}/D_{50}$ ratio could depend on the selected grain size percentile (e.g., the $D_{50}$ versus the $D_{90}$), which sets the transport threshold. However, recent experiments have shown that the $D_{90}$ better characterizes the thresholds for sediment entrainment than the $D_{50}$ (MacKenzie et al. 2018). Besides, grain size $D$ linearly propagates into the equation (2) and thus into the probability of $\tau > \tau_c$. Therefore, although the resulting probabilities vary depending on the threshold grain size, the relationships between the $D_{90}/D_{50}$ ratio and the mobilization probability will not change. In addition, because of the linear relationship between the $D_{90}/D_{50}$ and $D_{50}/D_{50}$ ratios (Fig. 2), the same dependency of transport probability on sorting will also emerge if the $D_{90}/D_{50}$ ratios are used. For the case where different discharge estimates are considered, here expressed as the ratio $\Delta$ of a specific runoff to the mean annual discharge $Q_{\text{mean}}$, then the corresponding probability of sediment transport will change by $\sim \sqrt{\Delta}$ (equation 7), but the dependency on the $D_{90}/D_{50}$ ratio will remain. This is also valid if transport probabilities are calculated based on discharge quantiles (Supplement S3), and if larger channel widths particularly for Peruvian streams are considered (Supplement S5).

This suggests that the sorting of the bed material has a measurable impact on the mobility of gravel bars and thus on the frequency of sediment mobilization irrespective of the selection of a threshold grain size and the choice of a reference water discharge. We note that while the data is relatively scarce and scattered (i.e., the same transport probability for a c. twofold difference in the $D_{90}/D_{50}$ ratio), the relationships observed between the probability of transport occurrence and the degree of material sorting are significant with p-values $<$0.01. We explain the scatter in the data by the natural stochastic nature of processes that are commonly encountered in the field.

For a given $D_{90}/D_{50}$ ratio, the probability of material transport tends to be greater in the Peruvian than in the Swiss rivers (Fig. 3A). We explain the albeit small divergence in the transport probability between both settings (i.e. regression parameters overlap within their 95% confidence interval) by (i) the differences between the geomorphic conditions and sediment supply processes in both mountain ranges, (ii) the anthropogenic corrections of the Swiss streams, and (iii) the generally stochastic nature in sediment supply. In the Swiss Alps, the channel network, the processes on the
hillslopes, and the pattern of erosion and sediment supply has mainly been conditioned, and thus controlled, by the glacial impact on the landscape and the large variability of exposed bedrock lithologies (e.g., Salcher et al., 2014; Stutenbecker et al., 2016). In addition, the intra-annual runoff variability is much lower than in the Peruvian streams (Supplement S2). In contrast, the erosion and sediment supply in the western Peruvian Andes is mainly driven by the combined effect of orographic rain (Montgomery et al., 2001; Viveen et al., 2019) and earthquakes (McPhilips et al., 2014), and the intra-annual runoff variability is quite large (Table 1). Because the patterns, conditions and mechanisms of sediment supply largely influence the grain size distribution of the supplied material (Attal et al., 2015), and as consequence, the downstream propagation of these grain size signals (Sklar et al., 2006), we do not expect identical relationships between grain size parameters and probability of sediment transport in both mountain ranges. In the same sense, the large variability and stochastic nature of sediment supply and transport processes could also explain the large spread in transport probability that we report for the Swiss and Peruvian streams. A large spread in transport probability was also inferred for mountainous rivers in the USA (Torizzo and Pitlick, 2004; Pfeiffer and Finnegan, 2018). We note, however, that Pfeiffer and Finnegan (2018) report lower transport probabilities that range between 8% and nearly 100% for the West Coast, 1% and 12% for the Rocky Mountains, and <10% for the Appalachian Mountains. In a broader sense, these authors considered the ratio between sediment supply and sediment transport capacity as criteria for the incipient motion of bedload material, which differs from the mobility criteria that we set in this paper. However, the largest difference stems from the values of the database. While the \( D_{50} \) of Pfeiffer and Finnegan (2018) has nearly the same size as the \( D_{84} \) reported here, their channel gradients tend to be 3 times lower. Because shear stress linearly depends on gradient (equation 7), then the probability where \( \tau > \tau_c \) will be directly and proportionally affected by this. Nevertheless, even if we would select a different channel gradient, the probability of material mobilization will go down (most likely linearly), but the dependency of the transport probability on the grain size sorting will remain. Accordingly, despite the large scatter in the dataset, the relationships between transport mobility and sorting is statistically significant with \( p \)-values \(<0.01 \), which suggests the sorting of coarse-grained bed sediments has a measurable impact on the mobility of the bedload material. We therefore conclude that besides the generally accepted controls including transport regime and sediment supply (Dade and Friend, 1998; Church, 2006), the sorting of the bed material represents an additional, yet important variable that influences the mobility of material on gravel bars. In addition, further research could possibly disclose a mechanism where sediment supply, material sorting and transport probability may be closely linked through a positive feedback.

Figure 1

A) Map showing the sites where grain size data has been measured in the Swiss Alps. The research sites are close to water gauging stations; B) map showing locations for which grain size and water discharge data is available in Peru (Litty et al., 2017).
Relationship between ratio of the $D_{96}/D_{50}$ and $D_{84}/D_{50}$, implying that the $D_{96}$ grain sizes of the Maggia gravel bars are too large if the $D_{50}$ is taken as reference and if the other gravel bars are considered.

A) Relationships between the probability of sediment transport occurrence and the $D_{96}/D_{50}$ ratio, which we use as proxy for the sorting of the gravel bar, in the Swiss and Peruvian rivers. B) Normalized residuals that are plotted against the sorting. The normalized residuals do not show any specific and significant patterns.

Table 1

Channel morphometry (width and gradient), grain size and water discharge measured at the research sites. The table also shows the results of the various calculations (critical shear stress $\tau_c$, shear stress $\tau$ of a flow with a mean annual runoff $Q_{med}$, and probability of sediment transport occurrence related to this flow).

Data availability

All data that have been used in this paper are listed in Table 1 and in the Supplement.

Author contributions

FS and RD designed the study. RD conducted the Monte Carlo Simulation. PG provided the grain size data in the Supplement. FS wrote the paper with input from RD and PG. All authors discussed the article.

Competing interests

The authors declare that they have no conflict of interest.

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**Note:** Data from Swiss Rivers is from the Swiss Federal Office for the Environment (FOEN). Data from Peruvian Rivers is from Littau et al. (2017) and Schlunegger et al. (2017).
Swiss & Peruvian rivers:
$R^2 = 0.91$
p-value $= 7E-18$

Maggia river (CH)

Figure 2
Figure 3

A

Swiss rivers:
$R^2 = 0.74$
p-value = $2E-4$

Peruvian rivers:
$R^2 = 0.33$
p-value = $4E-3$

B

Normalised residual

$+1\sigma$

$-1\sigma$

$+0$

$-0$

mean

$D_{96}/D_{50}$