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An experimental analysis of bed load transport in gravel-bed braided rivers with high grain Reynolds numbers

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ABSTRACT

Laboratory experiments were performed with nearly uniform fluvial gravel (D_{50}=9 mm, D_{10}=5 mm and D_{90}=13 mm) to analyse the relationship between stream power and bed load transport rate in gravel-bed braided rivers at high grain Reynolds numbers. The values of the unit-width dimensionless bed-load rate q_b^* and unit-width dimensionless stream power $\omega*$ were evaluated in equilibrium conditions based on ten different experimental runs. Then, they were plotted along with values obtained during particularly representative field studies documented in the literature, and a regression law was derived. For comparison, a regression analysis was performed using the data obtained from laboratory experiments characterized by smaller grain sizes and, therefore, referring to relatively low grain Reynolds numbers. A numerical integration of Exner's equation was performed to reconstruct the local and time-dependent functional dependence of q_b^* and $\omega*$. The results led to the following conclusions: 1) At equilibrium, the reach-averaged bed load transport rate is related to the reach-averaged stream power by different regression laws at high and low grain Reynolds numbers. Additionally, the transition from bed to suspended load transport is accelerated by low Re*, with the corresponding bed load discharge increasing with stream power at a lower, linear rate. 2) When tested against the gravel laboratory measurements, the high Re* power law derived in the present study performs considerably better than do previous formulas. 3) The longitudinal variability of the section-averaged equilibrium stream power is much more pronounced than that characterizing the bed load rate, at least for high Re*. Thus, the stream power and its local-scale heterogeneity seem to be directly responsible for transverse sediment re-distribution and, ultimately, for the determination of the spatial and temporal scales that characterize the gravel bedforms. 4) Finally, the stochastic interpretation of the wetted bed elevation function mapped in a particularly representative sample run reveals that gravel-bed transport and braiding are associated with the persistence of multiple equilibrium energy states due to the presence of truly non-stationary, local-scale pseudo-periodicity.

1. Introduction

Sediment transport in river flows constituted one of the main research topics of both hydraulic engineers and researchers working in the field of the fluid dynamics throughout the 20th century. Based on Bagnold, there were two main motivations leading to the formulation of a considerable number of controversial theories. The first motivation was that an inductive approach unavoidably involves approximations and practical difficulties when applied to very complex physical phenomena. The second was that the validation of the proposed formulas was carried out based on their immediate applicability to a specific problem rather than on their compatibility with field observation of the general underlying process. To fill the gap, Bagnold himself (1966), proposed a different and innovative interpretation of sediment transport in natural channels based on energy principles, providing a stepping stone for other researchers to evaluate the so-called critical stream power. This characteristic parameter is needed to identify the condition of incipient motion and characterize the different expected types of transport associated with different levels of flow intensity. Furthermore, it is crucial for the quantification of the sediment transport rate. The soundness and efficacy of Bagnold's approach were later proved by Gomez and Church (1989), who analysed the performances of the most popular bed load transport formulas at the time using field and laboratory data obtained in pseudo-steady flow conditions and available in the literature. They came to the conclusion that the energy (or stream power) approach should be preferred when estimating the sediment transport magnitude, mainly in the presence of limited hydraulic information as in the case of gravel-bed rivers, because it provides a straightforward correlation scale of the phenomenon. In its original version, Bagnold's

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NOTATION

A	bed cross-section area
B	total wetted width
B_{act}	active width
D	representative sediment diameter
e_b	bed load efficiency coefficient
ε_λ	wetted bed elevation experimental semivariogram
F_r	Froude number
g	acceleration of gravity
G_s	bed load mass transport rate
γ	specific weight of water
H	section-averaged flow depth
j	energy slope
λ	wetted bed elevation function
n	bed porosity
ν	water kinematic viscosity
Ω	stream power
$\omega*$	dimensionless unit-width stream power
P	total submerged weight transport rate
Q	water flow rate
q_b^*	dimensionless unit-width bed load transport rate
Q_b	bed load volume transport rate
Re^*	grain Reynolds number
R_H	hydraulic radius
ρ	water density
ρ_s	sediment density
S	bed slope
t	time
$\tan\alpha$	sediment shear coefficient
θ	Shields parameter
u^*	shear velocity
U	section-averaged velocity
x	longitudinal spatial coordinate
w_s	settling velocity
W	flume width
y	transverse spatial coordinate
z	bed elevation function

theory can be summarized by the following equation:

$$P = \Omega\left(\frac{e_b}{\tan\alpha} + 0.01\frac{U}{w_s}\right) \tag{1}$$

where P indicates the total submerged weight transport rate, $\Omega = \gamma Q j$ represents the so-called stream power, γ is the water specific weight, Q is the volume flow rate, j is the energy slope, $0.11 \le e_b \le 0.15$ is the bed load efficiency coefficient, $\tan\alpha$ is the sediment shear coefficient, U is the section averaged flow velocity and w_s is the settling velocity. The available stream energy represented by Ω may be replaced by its effective value, which is obtained by subtracting the critical threshold Ω_c from the total. This value is associated with the inception of the transport process. A few empirical expressions were proposed to compute that parameter based on the power law (e.g., Costa, 1983; Williams, 1983; Jacob, 2003; Gob et al., 2003)

$$\omega_c = aD^b \tag{2}$$

where $\omega_c = \Omega_c/B$ **(3)** is the unit-width critical stream power, B is the wetted width, D is the representative grain size, and a and b are two numerical coefficients ranging from 0.0253 to 0.03 and from 1.62 to 1.69, respectively.

The widely used Meyer-Peter's equation (1948), here re-written in unit-width dimensionless terms:

$$q_b^* = 8(\theta - \theta_c)^{3/2} = 8(\theta - 0.047)^{3/2} \tag{4}$$

was one of the first empirical formulas obtained by searching for a specific correlation between the intensity of the bed load transport and the intensity of the water flux.

In Eq. (4), q_b^* indicates the following dimensionless quantity:

$$q_b^* = \frac{Q_b}{B\sqrt{g\frac{\rho_s-\rho}{\rho}D^3}} \tag{5}$$

where Q_b is the total bed load transport rate, ρ and ρ_s are the water and the sediment densities, respectively, θ is the Shields parameter:

$$\theta = \frac{\gamma R_H j}{(\gamma_s - \gamma)D} = \frac{\Omega}{UB(\gamma_s - \gamma)D} \tag{6}$$

where $\gamma_s = g\rho_s$, and R_H is the hydraulic radius. The studies by Meyer-Peter and Müller (1948) inspired a considerable number of similar contributions, including

a) Hunziker (1995):

$$q_b^* = 5(\theta - 0.05)^{3/2} \tag{7}$$

and b) Wong and Parker (2006):

$$q_b^* = 4.93(\theta - 0.047)^{3/2} \tag{8}$$

More recently, Bertoldi et al., (2009) performed a series of laboratory tests based on Froude similitude and obtained a dimensionless equation relating unit-width bed load transport rate and unit-width stream power in the form of a monomial power law:

$$q_b^* = 0.412\omega^{*2.27} \tag{9}$$

where

$$\omega^* = \frac{\Omega}{\gamma B\sqrt{g\frac{\rho_s-\rho}{\rho}D^3}} \quad 0.15 \le \omega^* \le 0.45 \tag{10}$$

They also provided a diagram displaying the proportionality relationship between the ratio of active width B_{act} to wetted width B and $\omega*$. The term active width (e.g., Ashmore et al., 2011) refers to the portion of the cross-section top width where the shear stress is larger than the incipient motion threshold. Therefore, the ratio of that quantity to the wetted width as a function of the dimensionless stream power represents a measure of the expected degree of freedom associated with a given level of flow energy when modelling the cross-section geometry.

Most of previous laboratory experiments on sediment transport in gravel-bed braided rivers were performed at relatively low grain Reynolds numbers $\mathrm{Re}^* = u^*D/\nu$, with $u*$indicating the shear velocity and ν the water kinematic viscosity (e.g., Viparelli and Pica, 1967; Young and Davies, 1991; Zarn, 1997; Bertoldi et al., 2009). Their results were discussed in terms of reach-scale, equilibrium-state transport characteristics.

The aim of the present study was to analyse sediment transport in gravel-bed braided rivers at high grain Reynolds numbers, at reach and cross-sectional scale in both transient and equilibrium conditions. Note that “equilibrium conditions” mean average bed slope equal to average water surface slope, and sediment discharge and bar geometric features characterized by nearly constant values (Parker, 2003). For that purpose, new well-controlled laboratory experiments were performed at the University of Basilicata and large amounts of laboratory and field measurements already documented in the literature were processed for comparison. The reason for performing the experiments at relatively high grain Reynolds numbers was to reproduce turbulent and hydraulically rough flows in natural gravel-bed rivers (so that grain detachment and motion could be considered totally independent of the fluid viscosity) and to avoid turbidity and suspended loads. To investigate the reach-scale characteristics of gravel-bed sediment