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Assessment of bedload equations using data obtained with tracers in two coarse-bed mountain streams (Narcea River basin, NW Spain)

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ABSTRACT

This paper evaluates the predictive power of nine bedload equations, comparing the results provided by the equations with the bedload rates obtained in a previous field-based tracer experiment accomplished in River Pigüeña and River Coto, two coarse bed streams from NW Spain. Rivers from NW Spain draining the northern watershed of the Cantabrian Mountain range flow into the Bay of Biscay in a short path (50–60 km). In this region, they are developed forested catchments featured by fluvial networks with relatively steep slopes, single-thread sinuous channels, and where bed sediment is typically coarse (cobble and gravel).

Tagged stones were used to trace bed sediment movement during flood events in River Pigüeña and River Coto, the two main tributaries of the Narcea River basin. With the tracer results, bedload transport rates between 0.2 and 4.0 kg/s were estimated for six flood episodes.

The tracer-based bedload discharges were compared with the bedload rates estimated with the bedload formulae (DuBoys–Straub, Schoklitsch, Meyer Peter–Müller, Bagbold, Einstein, Parker–Klingeman–McLean, Parker–Klingeman, Parker and Wilcock-Crowe). Our assessment shows that all of the bedload equations tend to overestimate when compared with the tracer-based results, with the Wilcock and Crowe (2003) equation the only exception in River Pigüeña.

We linked these results to the particular geomorphology of coarse-bed rivers in humid and forested mountain environments. Within these rivers, armored textures and structural arrangements in the bed are ubiquitous; these features, together with a low sediment supply coming from upstream forested reaches, define a supply-limited condition for these channels limiting the potential use of bedload equations. The Wilcock and Crowe (2003) equation introduces complex corrections into the 'hiding function', and this could explain why it performs better.

1. Introduction

Bedload represents an important fraction of the total sediment load carried by the fluvial system. It controls channel morphology and dynamics, as well as extension of in-channel habitats (Dufour and Piègay, 2009). Consequently, fluvial research and management requires understanding of bedload dynamics. But estimation of bedload transport rates has been revealed as a very difficult task, particularly in coarse-bed rivers: under natural conditions bedload discharge is not a steady process, and it shows a strong variability; spatial and temporal (Batalla, 1997; Frostick and Jones, 2002).

Numerous sampling devices and field techniques have been developed in order to quantify bedload transport. Five principal ways of determining bedload discharge are described in the scientific literature: use of samplers (Helly and Smith, 1971; Sterling and Church, 2002; Vericat et al., 2006); installation of sediment traps on the channel (Laronne et al., 1992a, b; Reid et al., 1995; García et al., 2000; Bergman et al., 2007); the use of tagged clasts as 'bedload tracers' (Haschenburger, 1996; Haschenburger and Church, 1998; Hassan and Ergenzinger, 2003); a 'morphological' method, based on the quantification of geomorphological changes (Martin and Church, 1995; Ashmore and Church, 1998; Ham and Church, 2000; Fuller et al., 2003; Raven et al., 2010); and finally, new geophysical and acoustic methods (for example, Rickenmann, 1997; Rennie et al., 2002; Rennie and Villard, 2004; Belleudy et al., 2010).

The proper evaluation of bedload dynamics needs good records of bedload data, although obtaining long records is complex, expensive, and a time consuming task. The use of samplers and the tracer technique demand time-consuming field campaigns difficult to accomplish in the context of short-term river engineering projects. The morphological method requires time-series of topographical or photogrammetric measures, and this kind of data are not always available. Finally, geophysical methods are still incipient, so more research should be developed concerning how to address signal processing and/or calibration.

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Thus, for many practical purposes (for example, prediction and planning in the fluvial environment, river restoration projects) bedload transport is approached by using bedload formulae (López et al., 2013; Recking, 2013). The development of these formulae is based on the following premise: a functional relation should exist between the rates of bedload transport, the hydraulic variables, and the sedimentological conditions of the channel (Gomez and Church, 1989; Batalla, 1997; López et al., 2013).

Several bedload discharge formulae have been developed during the last decades, based mainly on laboratory data taken in flume and/or numerical modeling. Few of them were built using field data (Schoklitsch, 1950; Parker et al., 1982; Bathurst, 2007). Available bedload discharge formulae have been classified by Graf (1971), Gomez and Church (1989), and Habersack and Laronne (2002) into 'du Boys-type' equations (Du Boys, 1879) that have a shear stress relationship; 'Schoklitsch-type' equations (Schoklitsch, 1934) that have a discharge relationship; 'Einstein-type' equations (Einstein, 1950) that are based upon statistical consideration of lift forces; and 'Bagnold-type' stream power equations (for example, Bagnold, 1980).

The required input data (grain size, hydraulic conditions) for an adequate performance of bedload formulae are not always available in detail, and in many practical situations bedload equations are applied using width-averaged river characteristics (Recking, 2013). Consequently, evaluating the different equations developed for predicting bedload discharge in gravel-bed rivers and comparing its predictions with the bedload discharges measured in natural rivers is highly interesting and should be recognized as part of the calibration process of any conventional bedload transport modeling.

This is the main aim of the current paper, where the authors compare the results obtained with nine bedload formulae and the bedload transport rates measured using tracers in two coarse-bed mountain rivers belonging to the Narcea River basin (NW Spain). The tracer-based bedload rates measured were taken from Vázquez-Tarrío and Menéndez-Duarte (2014). The nine evaluated bedload formulae are: Du Boys–Straub (Du Boys, 1879; Straub, 1935), Schoklitsch (1934, 1950), Meyer Peter and Müller (1948), Einstein (1950), Bagnold (1980), Parker-Klingeman-MacLean (Parker et al., 1982), Parker and Klingeman (1982), Parker (1990), and Wilcock and Crowe (2003).

The Schoklitsch (1934, 1950) formula was chosen because it has been used before to approach bedload discharge in gravel-bed rivers (for example, Bathurst et al., 1987; D'Agostino and Lenzi, 1999). The same happens with Meyer Peter and Müller (1948), which is probably the most widely used bedload transport equation (Church and Hassan, 2005; Wong and Parker, 2006; de Linares, 2007). Einstein (1950), Bagnold (1980), Parker and Klingeman (1982), Parker-Klingeman-MacLean (Parker et al., 1982), Parker (1990), and Wilcock and Crowe (2003) formulae have a strong physical and experimental basis; this is the main rationale why we settled on these equations for the current assessment. Finally, the Du Boys–Straub equation (Du Boys, 1879; Straub, 1935) was chosen for historical reasons but also because it is still cited in some texts on river hydraulics (for example, Graf, 1971; Martínez Marín, 2001).

Previous attempts in order to evaluate bedload transport equations were made by other authors, but as Habersack and Laronne (2002) stated, in many cases they were based on data taken in flumes and/or on field data taken using samplers whose trap efficiencies were in the range of 40-60% (Carson and Griffiths, 1987; Gomez and Church, 1989; Chang, 1994; Reid et al., 1996; Batalla, 1997; Bravo-Espinosa et al., 2003).

Other assessments of transport equations using its own field data were made by García and Sala (1998), using its own measures in River Tordera with a Birbeck-type sampler (García et al., 1999). Habersack and Laronne (2002) evaluated several equations using field data taken with a Birbeck trap in the River Drau (Austria), an alpine tributary from the River Danube catchment. Martin (2003) and Martin and Ham (2005) evaluated several equations using morphological data in the Vedder River and the lower Fraser River (Canada), respectively. Recking (2010) made a detailed analysis of the performance of several bedload equations in mountain sand-gravel rivers, partially based on flume data. More recently, López et al. (2013) assessed several equations in River Ebro (Spain), which is a large and strongly regulated river that drains to the Mediterranean Sea.

In this work, the performance of these equations is evaluated in two coarse-bed mountain streams belonging to the Narcea River basin (NW Spain). Rivers from NW Spain, draining the northern Cantabrian watershed, are typically short and steep streams. Unlike most of the previous field-based assessments of bedload formulae, climatic conditions in these rivers are temperate and humid, and upland areas of river catchments are highly forested.

During the last decades, land use changes and human works (dams, embankments) are inducing geomorphological changes in these rivers (Fernández et al., 2006; Vázquez et al., 2012). Related to these land use changes (loss of cropping areas, afforestation), a slow geomorphological trend consisting in active channel narrowing, loss of active gravel bars, vegetal growing in old lateral gravel bars, and loss of anabranches has been generally described in rivers from this region during the twentieth century (Fernández et al., 2006; Fernández and Fernández, 2008; Fernández and Anadón, 2010; Vázquez et al., 2012). Moreover, according to Santos Alonso (2011), the current activity of debris flow processes in these drainage basins seems to be low when compared to other mountain settings.

These general geomorphological features suggest a low sediment input into the fluvial network during the twentieth century, and we think that this sediment-starved condition could be restricting the range of applicability of the most common bedload formulas. As long as most of these formulas deal with transport capacity, the low sediment input could involve the actual bedload rates are far below the potential capacity of transport. In this geomorphological setting, river channel bed is often featured by armored textures, clast arrangements, and bed structures that strongly influence clast entrainment and bedload transport rates (Church and Hassan, 2005; Hassan et al., 2008). If one particular bedload formula does not succeed in considering all these constraints, then it could be unreliable when applied to gravel-bed channels in humid, forested basins subjected to land use changes as those studied here.

Studies in sediment transport in this region have been scarce until recent times (Prego et al., 2008; Vázquez-Tarrío and Menéndez-Duarte, 2014). The specific objectives of this study are: (i) evaluate several bedload equations using field data; and (ii) increase the comprehension of bedload dynamics and prediction in mountain rivers placed in humid, temperate, and forested conditions — particularly, those fluvial systems draining the northern watershed of the Cantabrian Mountain range (NW Spain).

2. Regional setting

2.1. Study site

The Narcea River (catchment surface of 1800 km^2) is characterized by a short path (around 50–60 km) and high gradients with headwaters at roughly 2000 m draining to the Bay of Biscay (Atlantic Ocean). Climate is template and humid with annual precipitations of 1100 mm being distributed throughout the whole year. In summer, fluvial flows decrease to values of 10–20% of the winter flow and the fluvial regime is pluvial (Prego et al., 2008). Bedrock geology comprises Paleozoic sedimentary rocks, (including limestones, quartzites, sandstones, and shales) and Precambrian metamorphic slates in the headwaters. This basement was compressed during the Variscan orogeny, and later it was uplifted in relation to the Alpine tectonic realm (Álvarez-Marrón et al., 1997).

The current regional relief is abrupt, with incised deep river valleys, steeply dipping hillslopes (average values of slope around 20° and

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higher), and remaining glacial landforms that are widely preserved above 1500 m. The highest summits are reached on the southern divide, around 2200 m. Vegetal cover in the basin is dominated by deciduous forests (mostly beech and oak species forests) in shadow slopes and bush formations (heather and gorse species) in sunny slopes and elevated areas. Regarding agricultural and livestock uses, grassland formations are also common. In general, plant cover is continuous through the whole drainage basin, only being interrupted in some particularly frequent rocky slopes in the most elevated areas of the basin.

2.2. Bedload transport measurements: Studied reaches

In previous works (Vázquez-Tarrío, 2013; Vázquez-Tarrío and Menéndez-Duarte, 2014), bedload transport rates were estimated in two reaches from the Narcea River basin using tagged clasts (painted and with inserted magnets). The two studied reaches were selected in River Pigueña and River Coto, the two main tributaries of the Narcea River basin (Fig. 1B).

2.2.1. River Pigüeña

In River Pigüeña, the study section was chosen on a lateral gravel bar located in the lower part of the river basin, 1–2 km upstream from the confluence of River Pigüeña with the main channel of River Narcea (Fig. 2A). The surface of the catchment draining to this point is 400 km^2.

Mean annual discharge is 4.4 m^3/s, while the average minimum and maximum annual discharges are 1.1 and 9.5 m^3/s, respectively. Bankfull discharge is 70 m^3, and this discharge corresponds to a flood with a recurrence interval of 1.5 years.

Tracers were seeded on the gravel-bar surface and not in the channel in order to be able to work safely during the high water stages following major floods. The low water river channel has a width of 25 m in this

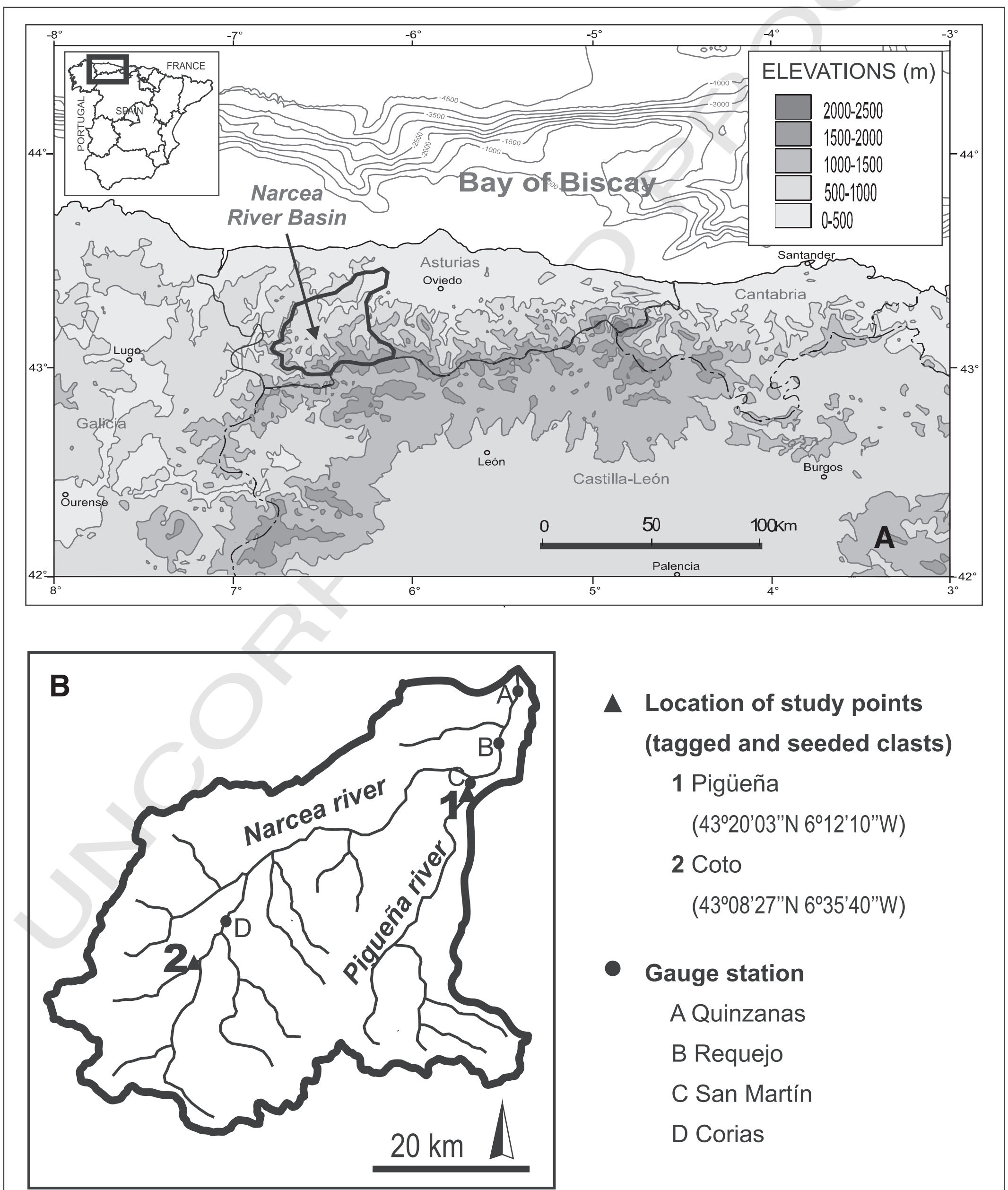

Fig. 1. (A) Location of River Narcea drainage basin in the northern Cantabrian Range watershed. (B) Location of the studied reaches (Rivers Pigüeña and Coto) and gauge stations along the River Narcea basin.

Fig. 2. (A) River Pigüeña lateral bar, (B) River Coto channel and lateral bar.

reach, and the bed slope measures 0.007. Following Montgomery and Buffington (1997), the channel could be classified as a pool–riffle channel in this reach. The gravel-bar width ranges between 20 and 90 m, and it has a length of, approximately, 90 m.

Bed material is represented mainly by cobble and gravel siliceous (quartzites) clasts. Surface D_{50} is 56 mm, while subsurface sediment is finer; subsurface D_{50} is 28 mm. Surface D_{50} is then two times coarser than subsurface D_{50}, suggesting a good degree of armoring.

The Folk and Ward (1957) sorting coefficient is 2.8, then the sediment is very poorly sorted (Bunte and Abt, 2001). The Sambrook Smith et al. (1997) bimodality index is 1.32, below the 1.7 threshold that Wilcock (1993) defined for bimodal sediment mixtures.

Sand fraction represents < 15% of the bulk sediment, being mainly concentrated in the spaces and openings between the subsurface sediment rather than in the surface-armored layer. Following Church (1978), bed state could be defined as 'underloose': bed surface is composed of close-packed and imbricated particles, with some microforms such as pebble clusters or boulder shadows.

2.2.2. River Coto

River Coto is placed in a more upstream position in the drainage network than River Pigüeña. In River Coto, the study section was also chosen including a lateral gravel bar (Fig. 2B), but in this case tracers were also seeded on the main channel. The surface of the catchment draining this point is 120 km^2. Bankfull discharge is 17 m^3/s, corresponding to the flow with a recurrence interval of 1.2 years.

The channel at base flow is 15 m at this point, and the bed slope is 0.01. Following Montgomery and Buffington (1997), the channel could be classified as a pool–riffle channel in River Coto. The gravel-bar width ranges between 10 and 15 m, and it has a length of, approximately, 60 m.

Bed material is composed of cobble and gravel siliceous clasts (mainly quartzites, but also metamorphic slates). Surface D_{50} is 88 mm and subsurface is 70 mm. Then, surface D_{50} is 1.2 times subsurface D_{50}, suggesting a mere faint degree of bed armoring, less conspicuous than in River Pigüeña.

The Folk and Ward (1957) sorting coefficient is 2, then the sediment is again poorly sorted (Bunte and Abt, 2001). The Sambrook Smith et al. (1997) bimodality index is 0.41: according to Wilcock (1993), bed sediment is unimodal.

Sand fraction represents < 6% of the bulk sediment, being mainly concentrated in the spaces and openings between the subsurface sediment rather than in the surface layer. Thus, bed state could be defined as 'underloose', the same as in River Pigüeña: bed surface is composed of close-packed and imbricated particles, with some microforms (pebble clusters).

3. Methodology

3.1. Previous tracer-based estimations

In River Pigüeña and in River Coto studied reaches, 1142 tagged clasts were seeded. During the hydrological years 2009–2010 and 2010–2011, six flood events with the ability to disturb tracer positions were analyzed — three of them occurring in River Pigüeña and three in River Coto. Table 1 summarizes the main features of the studied floods (date, time duration, discharge).

Water discharge was taken from the gauging records of the Puente San Martin and Corias gauge stations, close to River Pigüeña and River Coto studied reaches, respectively. Furthermore, minimum values of water depth for each transport episode were determined in the field looking for evidence (floating deposits, log deposits, water marks, etc.) of the water level reached by the flow (Fernández Iglesias, 2012). Then, knowing the water stage, the one-dimensional, mean boundary shear stress at each cross section was computed.

After these transport events, tracer displacements were measured along the main longitudinal direction of the channel. Following Eaton et al. (2008), they were tagged clasts belonging to the surface D_{50}

Table 1
Main features of the studied transport episodes (Vázquez-Tarrío and Menéndez-Duarte, 2014).

Date	River	Main peak time duration (h)	Maximum mean discharge (m^3/s)	Maximum peak discharge (m^3/s)	Basal shear stress (Pa)
15–18 January 2010	Pigüeña	72	32	104	115
10–24 June 2010	Pigüeña	102	80	100	112
31 October–20 November 2010	Pigüeña	43	79	108	118
13–16 Januray 2010	Coto	96	27	28	131
11–17 June 2010	Coto	47	28	30	135
6–8 January 2011	Coto	44	25	26	131

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semi-ϕ size class and the semi-ϕ size classes immediately upper and lower.

Based on Church and Hassan (1992), Laronne et al. (1992a, b), and Haschenburger and Church (1998), the measured displacements were used to estimate bedload transport rates: the average volumetric bedload transport discharge for each flood event was estimated using the product of the cross-sectional area of the moving mass of bedload and the average velocity of the bedload particles during the flow event.

In order to estimate the cross-sectional area of moving sediment two measurements are needed. On the one hand, for the active channel width, tagged clasts were seeded on the surface of the bed following a line perpendicular to the main flow direction; then disturbance of this line allows us to define the active width. On the other hand, for the active depth, we used two approaches: the depth of buried tracers and the active depth model suggested by Haschenburger (1999).

The average velocity of the bedload particles was derived from the displacements measured for the tracers and the Church and Hassan (1992) travel distance–grain diameter relation.

The disturbance of the tracer line also allowed us to observe the patterns of tracer dispersion and made some inferences about bedload transport conditions. In all the studied events, the whole length of the tracer line was disturbed. Moreover, clasts of all sizes were displaced and also remained stable; these two facts together suggest that bedload transport occurred in phase II or partial mobility conditions (Carling, 1988; Wilcock and McArdell, 1993, 1997) during the studied episodes: the condition in which all grain sizes are being moved, but only a portion of the grains on the surface of the bed ever move over the duration of a transport event (see Fig. 5 in Vázquez-Tarrío and Menéndez-Duarte, 2014).

Table 2 collects the bedload transport rates estimated for the studied transport events. Those bedload rates will be compared further in the text with the results obtained using bedload equations. Also, a good fit was found between the measured bedload transport rates and the one-dimensional, mean boundary shear stress at cross section. This regression equation will be used when evaluating the performance of the different bedload equations, and it follows the next expression:

$$q^{*} = 12.16 \cdot \left(\tau^{*} - 0.045\right)^{4.14}. \qquad (1)$$

In Vázquez-Tarrío and Menéndez-Duarte (2014), all the details about the measurement of the bedload transport rates and the tracer experiment are widely explained.

3.2. Selection and description of the bedload transport formulae

We selected nine bedload formulae that have been widely used in coarse-bed streams comparable to those studied here (Martínez Marín, 2001; Wilcock et al., 2009): Du Boys–Straub (Straub, 1935), Schoklitsch (1934, 1950), Meyer Peter and Müller (1948), Einstein (1950), Bagnold (1980), Parker–Klingeman–MacLean (Parker et al., 1982), Parker and Klingeman (1982), Parker (1990), and Wilcock and Crowe (2003).

Table 2
Bedload transport rates obtained with the tracer experiment in River Pigüeña and River Coto (Vázquez-Tarrío and Menéndez-Duarte, 2014); unit transport rate are the transport rates per unit of width of channel.

Flood episode	Studied channel	Measured transport rates (kg/s)	Unit transport rates (kg/m·s)
January 2010	River Pigüeña	4.06	0.10
June 2010	River Pigüeña	2.54	0.06
November 2010	River Pigüeña	1.10	0.03
January 2010	River Coto	0.20	0.01
June 2010	River Coto	0.21	0.01
January 2011	River Coto	0.28	0.01

The idea behind the development of those equations is that the intensity of bedload discharge is dependent on some hydraulic parameter that quantifies the magnitude of flow discharge; in general, they are functional relations of the following kind:

$$q = c \cdot \left(x - x_c\right)^{b} \qquad (2)$$

where q is the bedload transport rate, c and b are constant parameters determined empirically, and x is the parameter that represents flow discharge: discharge, shear stress or stream power.

As it was stated in the introduction, bedload equations could be classified in four main groups attending to the hydraulic parameter used to approach flow discharge (Graf, 1971; Gomez and Church, 1989; Habersack and Laronne, 2002).

In order to facilitate the exposition and the discussion of the results, in this work we have preferred to classify the different equations into two groups: 'classical equations' and 'modern equations'. Classical equations are those equations that are based on Eq. (2) (Du Boys, Shocklitsch, Meyer Peter-Müller, Einstein and Bagnold). Modern equations are based on the so called 'similarity collapse' hypothesis; they include complex considerations (Parker–Klingeman–MacLean, Parker–Klingeman, Parker and Wilcock–Crowe) concerning relative size effects, bed armoring, and the influence of sand content.

3.2.1. Classical equations

Du Boys–Straub (DB–S) represents the first proposed theoretical model for bedload transport (Du Boys, 1879; Straub, 1935). It is based on the 'excess shear stress' concept: the sediment transport will initiate once the basal shear stress in the channel reaches a threshold value. It was developed to describe the gravel motion in River Rhone (Gomez and Church, 1989), and it was used later by Straub (1935) in order to quantify the sediment transport in River Missouri.

The DB–S equation has been used in previous works, as for example Shulits and Hill (1968) and Gomez and Church (1989). Here, the Du Boys equation was used following the formulation suggested by Straub (1935):

$$q_b = k \cdot \tau \cdot \left(\tau - \tau_c\right) \qquad (3)$$

$$k = \frac{0.01003}{g \cdot D^{3/4}} \qquad (4)$$

$$\tau_c = \left(41.8 \cdot D^{0.82}\right) - \left[0.017 \cdot \ln(454 \cdot D)\right] \qquad (5)$$

where qb is the bedload rate, k is a coefficient depending on grain size D, τ is the shear stress and τc the critical shear stress for entrainment.

Schoklitsch (SC) (Schoklitsch, 1950) equation is based on discharge not on shear stresses. It was built using experimental data taken at the lab (Gomez and Church, 1989). In this work, this equation was used in the form proposed in Schoklitsch (1950):

$$q_b = 2500 \cdot S^{3/2} \cdot \left[Q - 0.6 \cdot \left(D^{3/2} / S^{7/6}\right)\right] \qquad (6)$$

where S is the cannel slope, Q is the water discharge per unit of channel width, and D the representative grain size for the bed sediment. The 40th percentile (D_{40}) of the grain size distribution has been used (Gomez and Church, 1989).

Meyer Peter and Müller (MP–M) equation (Meyer Peter and Müller, 1948) probably constitutes the most widely used equation when estimating bedload transport rates in natural rivers (Church and Hassan, 2005; de Linares, 2007). This equation was built based mainly on experimental data taken at the lab of ETH (Zurich, Switzerland), and it was initially based in flow discharge (García and Sala, 1998). Chien (1954) was able to express this equation in terms of the 'excess shear stress'. Years after, Chien's (1954) approach was improved by

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Wong and Parker (2006). In the current work the Meyer Peter–Müller formulae was used in the way suggested by Wong and Parker (2006):

$$q^{*} = 3.93 \cdot (\tau^{*} - 0.0495)^{1,5} \tag{7}$$

where q^{*} and τ^{*} are the dimensionless bedload rate and the dimensionless shear stress, which are defined by means of the Einstein (1950) and the Shields (1936) parameter respectively:

$$q^{*} = \frac{q_s}{\sqrt{(s-1)\cdot g \cdot D^3}} \tag{8}$$

$$\tau^{*} = \frac{\tau}{(s-1)\cdot \rho \cdot g \cdot D} \tag{9}$$

where s is the specific weight of sediment, g is the acceleration of gravity, and ρ is the density of the sediment.

The Einstein (EI) equation (Einstein, 1950) was one of the most complete and complex theoretical approaches to sediment transport (Graf, 1971; Gyr and Hoyer, 2006). It is a probabilistic model based on the stochastic nature of sediment entrainment. With his probabilistic approach, Einstein aimed to take into consideration the effects of turbulence and the effects of particle location in the bed (Gyr and Hoyer, 2006). The practical application of Einstein's (1950) model to real cases was very complicated (Gomez and Church, 1989; Martínez Marín, 2001), but Parker (1979) proposed an experimental fit based on shear stress. The latter is the approach followed in the current research:

$$q^{*} = 11,2 \cdot (\tau^{*})^{1,5} \cdot \left(1 - \frac{0.03}{\tau^{*}}\right)^{4,5}. \tag{10}$$

The Bagnold (BA) equation (Bagnold, 1980) is based on stream power, which Ralph Bagnold considered a useful parameter when quantifying the geomorphological work made by rivers on the landscape. The BA equation follows the next mathematical expression:

$$q_b = \frac{s}{s-1} \cdot 0.01 \cdot \left[\frac{\omega - \omega_0}{0.5}\right]^{3/2} \cdot \left({}^{Y}\!/_{0.1}\right)^{-2/3} \cdot \left({}^{D}\!/_{0.0011}\right)^{-1/2} \tag{11}$$

where s is the specific weight of the bed sediment, ω is the unit stream power, and ω_0 is the critical stream power. Stream power is estimated using the following function:

$$\omega = \frac{Q \cdot S \cdot \rho \cdot g}{B} \tag{12}$$

where B is the channel width, and S is the channel slope.

When computing ω_0, Ferguson (2005) introduced several corrections to Bagnold's (1980) formula. In this paper, computations were based on Bagnold's equation as suggested by Ferguson (2005):

$$\omega_0 = 2860 \cdot (D_{50})^{1.5} \cdot \log\left(0.235 \cdot \frac{D^{50}}{D_{90} \cdot S}\right) \tag{13}$$

where D_{50} and D_{90} are the 50th and 90th percentile of the grain size distribution, respectively.

3.2.2. Modern equations

The different equations classified here as 'modern equations' share a common background. Firstly, all of them are based on the 'similarity collapse' hypothesis (Parker and Klingeman, 1982): according to this hypothesis, the shape of the functional relation between bedload discharge and shear stress is not dependent on grain size (Parker and Klingeman, 1982; Wilcock et al., 2009). Secondly, these equations also consider that 'critical shear stress' (the shear stress needed for entrainment) should vary with the grain size following an exponential function called the 'hiding function' (Parker and Klingeman, 1982; Parker, 2008; Wilcock et al., 2009).

Despite these general considerations, each one of the modern equations has its own particularities. For example, the Parker, Klingeman and MacLean (P–K–MC) equation (Parker et al., 1982) was developed using field data taken in Oak Creek (Oregon, USA) and Elbowe River (Canada), and it is based on the 50th percentile (D_{50}) of the subsurface grain size distribution. The Parker and Klingeman (P–K) equation (Parker and Klingeman, 1982) is similar to P–K–MC, but it firstly calculates fractional transport rates for each size class and secondly summarizes for the whole sediment mixture.

The Parker (P) equation (Parker, 1990) was also based on the bedload discharge data from Oak Creek. It was developed from P–K–MC, but it uses surface grain size distribution in its computations. This equation excludes sand sediment, assuming that during those transport events with the ability to displace gravel sediment the sand should be carried as suspension load (Wilcock et al., 2009).

Finally, the Wilcock and Crowe (W–C) model (Wilcock and Crowe, 2003) was developed based on experiments carried out in flumes with mixed sand–gravel sediment. Sand is explicitly considered in this model based on Wilcock et al. (2001), who observed how in recirculating flumes sand sediment increases gravel mobility.

3.3. Performance, comparison and evaluation of bedload formulae

In this work, the different bedload equations were performed following different procedures. Firstly, the bedload discharges corresponding to each flood event recorded in the Pigüeña and Coto rivers were estimated based on tracers. Based on the hydraulic parameters (discharge, shear stress) associated with these large floods (Table 1), bedload discharges were also computed using the different equations in order to be compared with the tracer-based estimates.

In the case of the modern equations, the calculations were accomplished using BAGS, a PC-based software developed to compute bedload transport in gravel-bed streams (Pitlick et al., 2009; Wilcock et al., 2009).

After carrying out the flood event computation, the bedload rating curve was constructed for each equation. The bedload rating curves plot bedload transport rates as a function of shear stress.

The DB–S, MP–M, and EI equations are based on the shear stress, therefore these curves were built directly assigning values to shear stress in the equations. For modern equations, the bedload rating curve built with BAGS was used. BAGS uses in its computations an algorithm based on the Keulegan's resistance formulae and the Manning–Strickler equation (Pitlick et al., 2009; Wilcock et al., 2009). Furthermore, BAGS algorithms work with the partition between grain and form resistance when computing bed shear stress using the following formula, which is derived from the Manning–Strickler equation:

$$\tau' = 17 \cdot (S \cdot D_{65})^{1/4} \cdot U^{3/2} \tag{14}$$

where U is the flow velocity; it is calculated using the Keulegan resistance formula:

$$\frac{U}{\sqrt{g \cdot R \cdot S}} = 2.5 \cdot \mathrm{Ln}\left(11 \cdot {}^{R}\!/_{k_s}\right) \tag{15}$$

where R is the hydraulic radius, and k_s is the equivalent roughness that was calculated as two times the 65th percentile of the grain size distribution ($2D_{65}$).

The shear stress calculated using Eq. (14) is used by BAGS when performing the P–K–MC, P–K, P and W–C equations.

Finally, the SC and BA equations are based on discharge (stream power could be calculated from the discharge). We use the topographical channel section built with Total Station in order to compute the relation between hydraulic radius and the wetted perimeter of the

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channel cross section. Then, using the Keulegan equation the relation between hydraulic radius and the mean flow velocity was computed. Finally, based on the hydraulic radius–slope product for shear stress and the wetted perimeter–flow velocity product for discharge, the relation between shear stress and discharge was constructed for the two studied streams. Based on this relationship, it was possible to build the bedload rating curve.

The bedload discharges computed with the different equations were compared with the bedload rates obtained with the tracer experiment. To evaluate the performance of bedload rates, the comparison was made in two different ways:

- By means of a ‘discrepancy index’ (r), which could be defined as the ratio between the calculated (with equation) and the observed (with tracers) bedload rates.
  That index is similar to the one used by Batalla (1997) or Habersack and Laronne (2002) in their assessment of bedload transport equations. Following Habersack and Laronne (2002), the geometric mean of those indexes was also computed using the following expression:

$$\hat{r} = (r_1 \cdot r_2 \cdot \ldots \cdot r_n)^{1/n} \tag{16}$$

  where n is the number of data.
- By comparing the bedload rating curve built using Eq. (1) with the rating curve built for each equation.

4. Results

In Tables 3 and 4 the results obtained with each equation are summarized and also the discrepancy indexes obtained when comparing the bedload rates calculated with each equation with the bedload rates obtained through the tracer experiment. Bedload rates estimated with the equations are, in general, higher than the bedload rates measured with tracers. The bigger discrepancies are obtained in River Coto; but in River Pigüeña differences are also important, particularly when performing the classical equations.

The same statistical indexes applied for the set of bedload formulae, were obtained for Eq. (1), and they are also included in Tables 3 and 4. This equation represents a regression fit derived from our own data in the study reach, and as such it is not comparable with the rest of formulae. Despite this, those indexes were calculated in order to facilitate the comparison between the equations and the bedload rates determined with the tracer experiment.

Fig. 3 shows the comparison between the bedload rates estimated with the different equations and the bedload rates obtained with tracers. Only 4% of our estimations are in range of 2 of the tracer measures, and only 13% are in a range of 10. The P–K–MC model represents the equation that provides better results for River Coto, while W–C is the equation that provides the better results for River Pigüeña.

The DB–S equation is the one with a higher discrepancy index (close to 20,000 in River Coto). The SC, MP–M, and EI discrepancy indexes are also high. In River Pigüeña, bedload rates obtained with P–K, P, and W–C are closer to the bedload rates obtained with tracers, being the discrepancy index lower for the January and June (2010) transport events. In the case of the W–C equation, the discrepancy index is close to 1 for these two transport episodes, which means that bedload rates estimated with the W–C equation and rates measured with tracers are almost the same. In River Coto, discrepancy indexes are high, even with the modern equations: the lowest discrepancy index (12.5) was obtained with the BA equation.

In Figs. 4 and 5 the rating curves built with each equation are compared with the rating curves derived from Eq. (1) for River Pigüeña and River Coto, respectively. Several equations overestimate or overpredict bedload rates for all the range of shear stresses: DB–S, EI, BA, and P. In the case of the P equation in River Pigüeña, both curves are very close at shear stresses around 50–70 Pa, which corresponds to frequent floods. On the other hand, the SC, MP–M, and P–K–M equations underestimate bedload rates with low shear stresses and strongly overestimate with moderate and high shear stresses. Finally, in the case of the W–C equation, both rating curves are very close in River Pigüeña, while in River Coto the W–C curve strongly differs from the experimental rating curve.

5. Discussion

Comparison of bedload discharges computed using the bedload equations with the tracer-based bedload rates measured shows systematically an overestimation of bedload transport rates with the equations, with the only exception being the W–C equation in River Pigüeña.

Our results show how the results given by the different equations are far from those obtained with tracers. Moreover, the results strongly differ when comparing the different equations one with each other. Classical equations give bedload discharges that are strongly higher than the bedload rates obtained with tracers. On the other hand, the modern equations give results closer to the transport rates obtained with tracers, although they provide still higher transport rates.

In principle, we could interpret these discrepancies in two different ways. On the one hand, we could attribute the discrepancies to the lack of reliability in the estimations made by the equations. On the other hand, the discrepancies could be attributed to inaccuracies or uncertainties in the measurement of bedload rates with the tracer experiment.

5.1. Tracer-based bedload rates

As noted in Section 3.1, tracer-based bedload rates were estimated here using the product of the cross-sectional area of the moving mass of bedload and the average velocity of the bedload particles during the

Table 3
Summary of the results obtained comparing bedload transport rates estimated using the equations, and the bedload rates obtained in the field with tracers; results for River Pigüeña.

	Estimated bedload rates (kg/s)			Discrepancy Ratios			
	January 2010	June 2010	November 2010	January 2010	June 2010	November 2010	Geometric mean
Du Boys–Straub	4526.4	4312.4	4753.9	1116	1696	4337	**2017.2**
Schoklitsch	316.4	298.4	335.1	78	117	306	**140.9**
Meyer Peter–Müller	481.0	453.5	509.4	118	178	464	**213.6**
Einstein	845.6	789.1	904.7	208	310	824	**376.3**
Bagnold	88.5	83.6	94.3	22	33	86	**39.5**
Parker–Klingeman–MacLean	100.2	91.1	110.1	25	36	101	**44.6**
Parker–Klingeman	32.6	26.6	39.9	8	11	36	**14.5**
Parker	18.7	15.8	22.2	5	6	20	**8.3**
Wilcock–Crowe	3.6	3.0	4.3	1	1	4	**1.6**
Eq. (1)	2.2	1.8	2.5	1	1	2.3	**0.9**
Tracer-based	4.1	2.5	1.1	–	–	–	–

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Table 4
Summary of the results obtained comparing bedload transport rates estimated using the equations and the bedload rates obtained in the field with tracers; results for River Coto.

Estimated bedload rates (kg/s)				Discrepancy ratios			
	January 2010	June 2010	January 2011	January 2010	June 2010	January 2011	Geometric mean
Du Boys–Straub	4101.1	4449.9	4101.1	20,725	20,876	14,907	**18,614.2**
Schoklitsch	544.1	585.4	541.0	2749	2746	1966	**2457.8**
Meyer Peter–Müller	647.6	697.5	643.9	3272	3273	2340	**2926.3**
Einstein	294.4	331.6	291.7	1488	1556	1060	**1349.0**
Bagnold	18.4	21.7	12.8	93	102	47	**76.1**
Parker–Klingeman–MacLean	2.4	4.1	2.3	12	19	9	**12.5**
Parker–Klingeman	3.9	6.2	3.7	20	19	14	**17.2**
Parker	63.9	78.3	62.9	323	367	229	**300.4**
Wilcock–Crowe	10.1	12.6	10.0	51	59	36	**47.8**
Eq. (1)	0.2	0.3	0.2	1	1	1	**1.0**
Tracer-based	0.2	0.2	0.3				

flow event. Then, reliability in the results depends in some way on the criteria followed when defining the cross-sectional area of the moving mass of the bedload particles and the virtual velocity of bedload.

Cross-sectional area of the moving mass of bedload depends on active channel width and scour depth. Both constraints should be discussed. Tracers were seeded following transversal lines that were disorganized in practically its whole length during all studied flow events. It seems reasonable to assume that the whole channel width was active during at least one moment through the course of the studied transport episodes. Consequently, channel bed perimeter measured over the channel cross section was used as active channel width.

Concerning the active depth, we found active depth values of roughly 20–25 cm in River Pigüeña and around 7–7.5 cm in River Coto using the scour-and-fill depth model developed by Haschenburger (1999). Those represent values around D_{90} in River Pigüeña and around D_{50} in River Coto and are in the same order of magnitude of the depth measured for the recovered buried tracers (15 cm in River Pigüeña; 5–10 cm in River Coto).

Virtual velocity of sediment was determined from the measured tracer travel distances, which according to Haschenburger and Church (1998) is a reliable strategy. Strictly speaking, our virtual velocities applied to 63% of channel sediment in River Pigüeña and 72% in River Coto because particles < 16 mm and > 256 mm were not monitored. Nevertheless, Church and Hassan (1992) have shown that the sensitivity of travel distance to particle size lessens as size decreases below the median diameter of subsurface sediment. For particle sizes above the median diameter, Church and Hassan (1992) proposed a heuristic model based on the travel distance–grain diameter relation; this model was used here for the coarser particles. This model has been confirmed by Haschenburger (1996) in the field and by Wilcock (1997) experimentally.

Then, while some uncertainties in the bedload rate estimations using the tracer-based method are unavoidable, the previous paragraphs show that the major potential sources of error were controlled with field data and no strong discrepancies should be expected with the actual rates.

After using the tracer method to estimate bedload rates and following a similar workflow, Liébault and Laronne (2008) estimated bedload yields very close to the volumes measured with sediment trapping: they found a 12% underestimation of bedload yields with the tracer method.

In our case, there is not an external source of data to use for assessing our estimations, but the bedload rates measured are in the same order of magnitude as bedload rates obtained in comparable coarse-bed streams from other regions of the world (Haschenburger and Church, 1998; D'Agostino and Lenzi, 1999; Batalla et al., 2005a, b).

Instead, the bedload discharges reported by the equations are two or three orders in magnitude higher than those bedload rates obtained with tracers, and at the same time the different equations show important discrepancies between them. Thus, while it is true that some uncertainty in the exact value of the bedload rates is unavoidable, differences between tracer-based estimations and results from bedload equations are still very high.

Even assuming in our estimations the 12% underestimation found by Liébault and Laronne (2008), our tracer-based bedload rates still would be far below the bedload rates estimated with the equations. In addition, inaccuracies in our tracer-based bedload rates would not explain the significant differences between the results provided by the different equations. Therefore, it is reasonable to mainly attribute the discrepancies between tracer-based bedload rates and bedload equation estimates to the lack of reliability in the estimations made by the equations.

5.2. Reliability of equations

With the results obtained here, the DB–S equation does not seem to be useful for the studied channels. According to Gomez and Church (1989) and Martínez Marín (2001), that equation was developed using finer sizes, and it is based on a very simplistic model of sediment transport based on the sliding of several layers of sediment within the river bed. These conditions could not be assumable in coarse-bed streams as those studied here.

In some works, good results have been reported for the SC equation (García and Sala, 1998; D'Agostino and Lenzi, 1999). However, in the current research it does not seem to provide good estimations. The same could be said about the MP–M equation.

The EI equation shows results similar to the previous formulas. The EI equation was developed based on flume data taken with sediment finer than the studied channel and that fact should be taken into account. The grain size of the bed sediment in the studied reaches probably is very coarse to be used with EI equation.

Regarding the BA equation, previous authors have found that this equation, as well as other mathematical models founded on the stream power concept, do better predictions of bedload transport rates (Gomez and Church, 1989; Martin and Church, 2000; Martin, 2003). Nevertheless, the differences between bedload discharges computed in this work using the BA equation and the tracer-based bedload rates are still important. So, despite the BA equation providing better estimations than the other classical equations, it did not give us good enough results.

In general, the modern equations provided better estimations. From the modern equations, in River Pigüeña the P and W–C equations showed the lowest discrepancy indexes, with the W–C equation providing better results than the P model. The W–C and P models are surface-based models, while the P–K–MC and P–K are subsurface-based models. Also, the P–K model, which is a multifraction model, provided better estimations than the P–K–MC model, which is based on subsurface D_{50}. Therefore, lower discrepancy indexes obtained with the P and W–C models in River Pigüeña could be suggesting that surface-based models are more reliable in this river reach than subsurface-based

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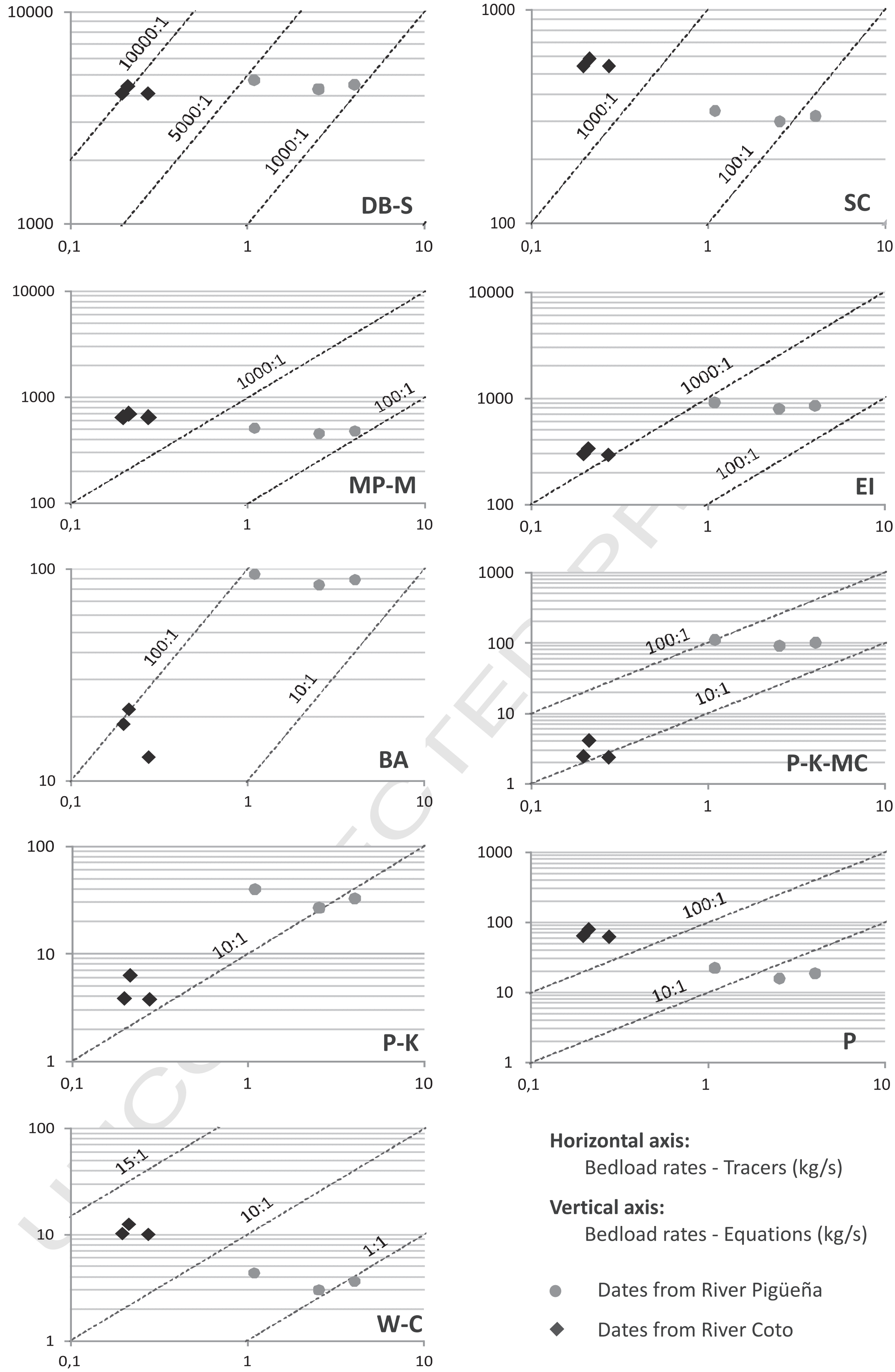

Fig. 3. Comparison between bedload rates estimated through the different equations (*y*-axis), and the bedload rates derived from tracer experiment (*x*-axis). The nine plots show the results obtained with each one of the nine bedload equations assessed here: Du Boys–Straub (DB–S); Schoklitsch (SC); Meyer Peter–Müller (MP–M); Einstein (EI); Bagnold (BA); Parker–Klingemann–McLean (P–K–MC); Parker–Klingemann (P–K); Parker (P); Wilcock–Crowe (W–C).

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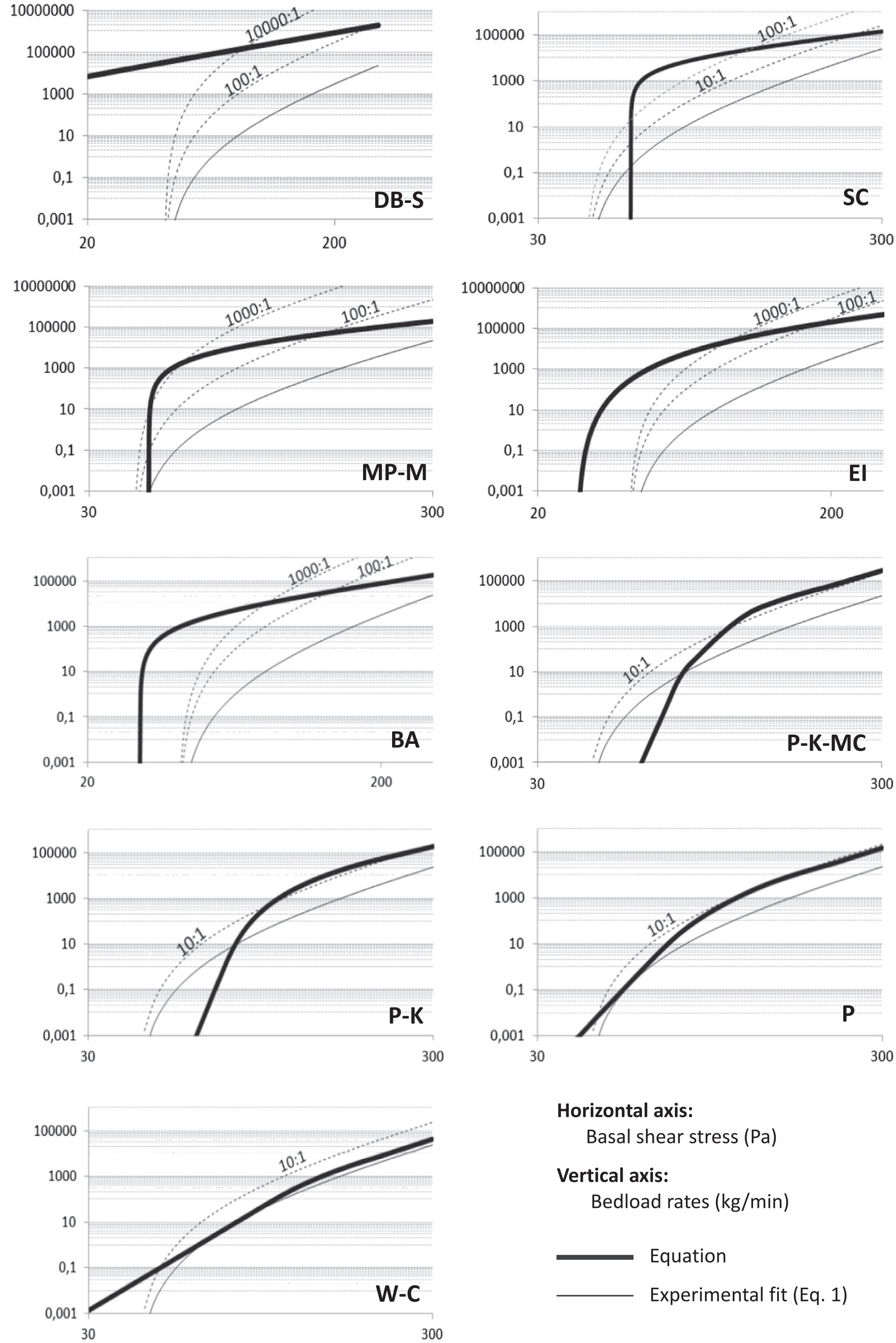

Fig. 4. Bedload rating curve built using each theoretical equation, compared with the bedload rating curve derived from the experimental fit (River Pigüeña): Du Boys–Straub (DB–S); Schoklitsch (S); Meyer Peter–Müller (MP–M); Einstein (EI); Bagnold (BA); Parker–Klingemann–McLean (P–K–MC); Parker–Klingemann (P–K); Parker (P); Wilcock–Crowe (W–C). Dotted lines show the values that represent a ratio of 10, 100, 1000…times the results obtained with the experimental fit.

models, at least at event scale. This also has been suggested in previous works, like Parker (1990), Wilcock and Crowe (2003), de Linares (2007), and de Linares and Belleudy (2007).

The W–C model provides better results than the P equations, in River Pigüeña and in River Coto. When developing the P model, Parker (1990) did not consider the sand sediment content, while Wilcock and Crowe

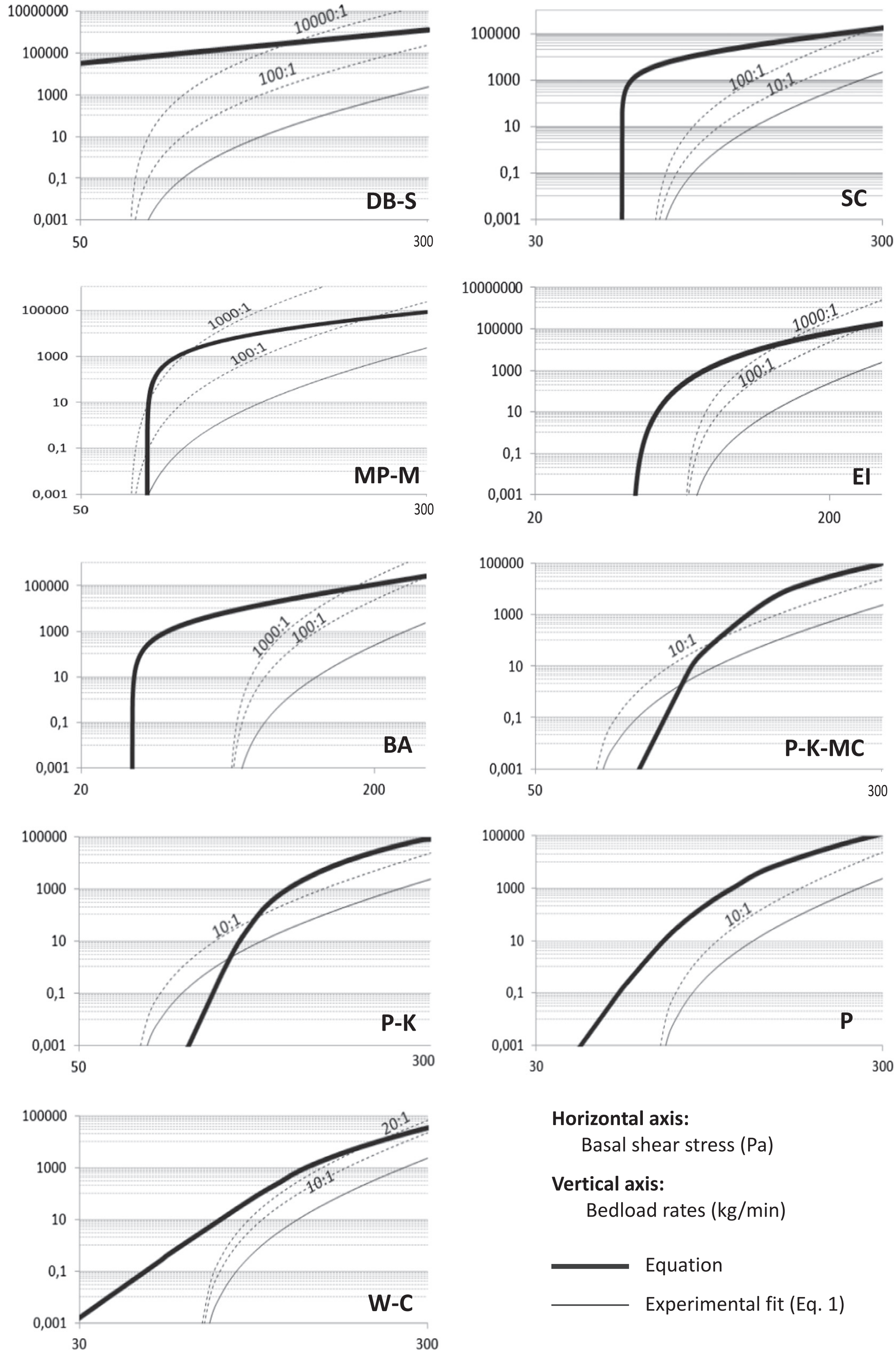

Fig. 5. Bedload rating curve built using each theoretical equation compared with the bedload rating curve derived from the experimental fit (River Coto): Du Boys–Straub (DB–S); Schoklitsch (S); Meyer Peter–Müller (MP–M); Einstein (EI); Bagnold (BA); Parker–Klingemann–McLean (P–K–MC); Parker–Klingemann (P–K); Parker (P); Wilcock–Crowe (W–C). Dotted lines show the values that represent a ratio of 10, 100, 1000…times the results obtained with the experimental fit.

(2003) tried to account for the nonlinear effects of sand content on gravel transport rates. At first sight, we could think that this explained why the W–C model performed better than the P model. But as it was stated above, those are surface-based models and the sand content in the surface sediment reaches is almost zero in both studied reaches. So our results suggest that the W–C performs better than the P model even in the absence of sand-content in the surface sediment.

Moreover, the W–C gives very good results in River Pigüeña but not in River Coto where it still overestimated the bedload rates by one order of magnitude. The same is observed with the P model, where discrepancies are higher in River Coto than in River Pigüeña. Furthermore, in River Coto the P–K–MC and P–K subsurface-based models provided better estimates than the W–C and P models.

We think that this different behavior between River Pigüeña and River Coto could be linked to the differences in the grain size distributions: median grain size is finer in River Pigüeña, and sediment sorting is slightly lower in River Coto. Also, armor ratio and bimodality index are lower in River Coto (1.2 and 0.41, respectively) than in River Pigüeña (2 and 1.3, respectively). These results could be suggesting some sensitivity to bed sediment texture when choosing between surface-based and subsurface-based models in the studied streams.

The W–C model deserves a deeper analysis because it provides very good estimations in River Pigüeña, although in River Coto strongly overestimates. The W–C model represents a surface-based multifraction model, with some degree of size selectiveness for sand and gravel sediment (de Linares, 2007; de Linares and Belleudy, 2007).

According to de Linares (2007), in some cases the W–C model does not approach well the actual degree of size selectiveness of bed sediment. The W–C model was developed to address mixed sand/gravel sediments. Size selectiveness of gravel increases with sand content (Wilcock and Kenworthy, 2002), so in the case of a sand/gravel sediment mixture de Linares and Belleudy (2007) found how the W–C model makes a correct estimation of threshold stresses. But when the bed sediment is dominated by a unimodal gravel mixture of gravel particles, gravel fractions have the same critical shear stress; in this particular case, de Linares and Belleudy (2007) found how the W–C model is not able to catch the actual degree of size selectiveness.

Furthermore, in natural streams in partial transport conditions, size effects on fractional transport rates are more noticeable for the coarser grain sizes (Church and Hassan, 2002). Therefore, as long as bed sediment is coarser in River Coto than in River Pigüeña, differences between estimated and actual bedload rates could be enlarged in River Coto because of the ineffectiveness of the W–C model to catch the actual fractional transport rates for the coarser particles.

In addition, this could explain why the P–K–MC model is the model providing better results in River Coto: the P–K–MC made an average estimation of bedload rates using the median size of sediment rather than doing a multifraction estimation and then summarizing for the whole sediment mixture; thus, it avoids the problem of catching the actual degree of size-selectiveness in the sediment mixture.

5.3. Comparison with previous field-based assessments

Discrepancies between the estimates made with bedload equations and the real measurements of bedload rates (using either field or flume data) were also found by other authors. Gomez and Church (1989) in an analysis made over 12 bedload equations developed for gravel-bed rivers found that any equation was able to do general predictions of bedload transport rates.

Reid et al. (1996) found that the MP–M equations gave better results than the BA and P in an ephemeral gravel-bed stream from Israel (Nahal Yatir). According to García and Sala (1998), in the latter case the river has not surface armoring; thus sediment availability was not conditioned by armoring, and sediment mobility was only controlled by stream capacity. Conversely, García and Sala (1998) observed how, with bed armoring, the P–K–M model is the one that provided better results using its own data obtained in River Tordera (NE Spain).

Hoey and Sutherland (1991) evaluated the BA equation and found that in equilibrium or aggrading rivers, this equation overestimated bedload rates; otherwise, in degrading channels this equation underestimated bedload rates. Nevertheless, in the channels studied here there is no evidence of aggradation, and the equation still overestimates the bedload transport rates.

Rascher et al. (2012) assessed 16 bedload equations in a mountain river in Bavaria, using bedload rates measured with a Helley–Smith sampler and obtaining a general overestimation of bedload rates. Furthermore, they observed how some of the evaluated equations overestimate during moderate to high flow while underestimating during low flows. This is similar to what Recking (2010) found using field and flume data.

More recently, López et al. (2013) evaluated various bedload equations using data obtained from natural and flushing floods studied during 2003 and 2004 in the Ebro River, a large regulated river draining to the Mediterranean Sea. They found in many cases discrepancy indexes of 2 to 10 between bedload equations and measured rates.

Our average percentage of bedload estimations in a range of 2, 5 and 10 the observed rates are in general lower than the percentages obtained by these previous studies (Table 5). However, our results are very close to those obtained by Rascher et al. (2012) in a comparable alpine stream. We think that our low percentages are partially related to the

Table 5
Performance of the formulae compared with a selection of recent studies in gravel-bed streams.

Reference	N[a]	r (0.5–2)[b] %	r (0.2–5)[c] %	r (0.1–10)[d] %	Observations
Batalla (1997)[e]	5	50	–	–	Bed-material load in a sandy, gravel-bed river.
Habersack and Laronne (2002) [e]	13	36	–	–	Alpine gravel-bed river
Martin (2003) [f]	4	19	44	75	Annual transport in 10 reaches of a gravel-bed river
Martin and Ham (2005) [e, f]	3	11	25	47	Average annual transport in 13 reaches of a gravel-bed stream
Duan et al. (2006)[e]	3	–	–	57	Low flow in two reaches of a desert gravel-bed stream
Recking (2010)[g]	4	13	27	34	6319 data from 84 reaches of sand- and gravel-bed rivers, and flume experiments
Rascher et al. (2012)[e]	16	–	–	19	Bedload rates in a coarse-bed river from Bavaria
Rascher et al. (2012)[h]	7	–	–	25	Bedload rates in a coarse-bed river from Bavaria
López et al. (2014) [e]	10	19	41	57	Regulated river experiencing cycles of armoring
This study	9	4	7	13	Bedload rates for flows in two mountain coarse-bed streams
This study[h]	4	8	17	29	Bedload rates for flows in two mountain coarse-bed streams

[a] Number of formulas tested in the paper.
[b] 0.5 < r < 2, the average percentage of predicted bedload discharge not exceeding a factor of 2 in relation to the observed discharge.
[c] 0.2 < r < 5, the average percentage of predicted bedload discharge not exceeding a factor of 5 in relation to the observed discharge.
[d] 0.1 < r < 10, the average percentage of predicted bedload discharge not exceeding a factor of 10 in relation to the observed discharge.
[e] Measurements with a Helley-Smith/basket sampler.
[f] Annual bedload yields obtained using the morphological approach.
[g] Assessment accomplished using field data taken from the scientific literature, and their own flume results.
[h] The average percentages are calculated only for the modern equations, not for all the equations assessed in the study.

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fact that some of the classical equations (DB–S, SC, EI, BA) evaluated here were not assessed by the previously cited authors. In fact, when we do not consider those classical equations, our percentage of agreement increases; the same happens with Rascher et al. (2012) results (Table 5). But even not considering classical equations, discrepancy indexes from Rascher et al. (2012) and from the present study are still lower than the other assessments. Therefore, we think it could be related to the particular geomorphological setting of the rivers studied here and by Rascher et al. (2012), differentiating them from the other studies: perennial streams with steep slopes, very coarse beds (gravel and cobble), and coming from mountain forested catchments in humid, rainy conditions.

5.4. Implications

5.4.1. Applicability of bedload equations in coarse-bed rivers

Recking et al. (2012) outlined the following facts in relation to the applicability of theoretical equations when used to estimate bedload transport rates in coarse-bed rivers: (i) There is no theoretical model with ability to make general predictions of bedload transport rates in gravel-bed streams (Gomez and Church, 1989); indeed, theoretical models are only acceptable in the short range of conditions for what they were defined. (ii) During low to medium magnitude flow conditions, equations show the worst results (Barry et al., 2004; Rascher et al., 2012). (iii) In coarse-bed rivers, errors could attain several orders of magnitude (García and Sala, 1998; Barry et al., 2004; Bathurst, 2007; Rascher et al., 2012). Furthermore, different authors got different conclusions because the range of data used by every author was different (Recking et al., 2012).

We believe that the previous three observations are applicable to our data. We did not find any model with the ability to do satisfactory predictions of bedload rates. At first glance, we think that there could be three main reasons explaining the strong overestimation found here: (i) we performed these equations averaging the flow conditions for the whole channel section and the whole duration of the transport episode, which involves assuming steady and uniform flow conditions; (ii) limitations in the definition of the parameters requested by the equations; and (iii) limitations inherent to the equations.

Firstly, in relation to point (i), Gomez and Church (1989), based in De Vries (1973) and Csoma (1973), found that a realistic comparison could be made if bedload rates estimated using bedload formulas were compared with the measurement of an average bedload rate that absorbs all the uncertainties linked to short-term fluctuations. That would be possible if the number of samples and the time interval of sampling are large enough to cover all the range of fluctuations (temporal and spatial) in transport rates.

We think that the use of tracers satisfies these requirements. Instantaneous bedload velocities or local bedload rates are not measured with tagged clasts. Rather than this, this technique allows us to determine average velocities of bedload (the virtual velocity of Hassan et al., 1992), based on data obtained with tracers dispersed across a wide surface of the channel section. In this sense, the observed transport rates are averaging bedload transport during the whole transport episode and across the whole width of the channel section. Following Gomez and Church (1989), we could consider that tracers are masking the effect caused by the fluctuations and unsteady behavior of bedload transport, allowing us to obtain average bedload rates. Thus, we think that the main discrepancy found in this work is not related to point (i).

In relation to point (ii), Recking et al. (2012) outlined several sources of uncertainty when applying the bedload equations: slope should be energy slope and not average bed slope (Meirovich et al., 1998); discharge should be measured locally, not using average values; and finally, grain size distribution should be properly measured.

According to these authors, all these facts result in the accuracy dependence of the time interval considered when performing the equations: when bedload equations are used to estimate bedload rates for very short time scales (instantaneous bedload rates), uncertainty is huge; however, the uncertainty decreases when the equations are used to calculate sediment transferences at longer time scales (for example, annual loads). They linked this to the fact that, at longer time scales, temporal fluctuations in the different parameters (slope, discharge, grain size) are averaged.

Therefore, following what was pointed out in the previous paragraphs, we could consider that tracer-based results averaged the temporal and spatial fluctuations that occurred during the transport episode, at least partially. Furthermore, as Habersack and Laronne (2002) stated, although the formulas theoretically require local rather than average cross-sectional hydraulic data (Gomez and Church 1989), for the derivation the originators of the equations used average cross-sectional data (Recking, 2013), straightforwardly available in practical situations.

On the other hand, when talking about the representative grain size introduced in the equations, Bravo-Espinosa et al. (2003) argued that transport conditions vary between the different grain sizes. Thus, they stated that estimating bedload transport rates using a unique grain size to represent the whole bed sediment mixture is not suitable. These authors defined three transport conditions for the different grain sizes of the bed sediment: those grain sizes that show supply-limited transport; those that show capacity-limited transport; and finally, those grain sizes that show partial capacity-limited transport. Then they observed how some equations seem to be more appropriate for each transport condition. For example, they observed how in capacity-limited conditions the P–K–MC (Parker et al., 1982) equation shows better results. This fact could be contributing in some way to the discrepancies observed here.

Finally, in relation to point (iii), we should highlight how using bedload equations in order to predict bedload rates involves assuming at least two tacit premises. Firstly, the application of a bedload formula implies assuming that during the transport event, not only flow conditions but also bed material and bedload sediment remain without changes: equations describe bedload as a steady process (Batalla, 1997). On the other hand, bedload formulas assume capacity-limited conditions: they compute the maximum load that the river channel is able to carry, and they do not take into account possible limitations in sediment supply that are common in natural systems (Wilcock et al., 2009; Recking, 2012).

The first assumption (steady bedload transport) is not realistic when talking about natural rivers. Not only because of the spatial and temporal fluctuations in flow conditions, but also in relation to the way bedload transport actually takes place. At event scale, bedload transport shows pulses (Gomez, 1991; Frostick and Jones, 2002) linked to the migration of bedforms and clusters of particles (Whiting et al., 1988). Also, transference of clasts from bed material to bedload follows a stochastic behavior (Kirchner et al., 1990). Moreover, at each particular moment during a transport episode, not all the water discharge is available for the transport (Gomez and Church, 1989). Furthermore, not all the bed surface is being involved in the active transport at every moment during the course of a transport episode; rather than this, in every moment different portions of the bed could be activated or inactivated, in relation to the evolution experienced by the texture of the bed, the structural arrangements, and the grain size of the bedload.

All these facts impose a chaotic and nonlinear nature to the bedload dynamics at event scale, which seems very difficult to be considered or included into a deterministic single equation. Flume-derived experiments from Recking (2013) lead to the conclusion that nonlinearity effects in bedload prediction are considerable for low transport stages, being mostly sensitive to the strong variance in shear stress at low flow.

In relation to the capacity-limited transport assumption, we think that this is one of the main reasons of the discrepancies found in the current research. We consider that bedload equations estimate not real bedload rates but transport capacity (Bravo-Espinosa et al., 2003; Wilcock et al., 2009).

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However, in coarse-bed rivers transport capacity could not be attained if there is not enough availability of sediment in the channel (Hicks and Gomez, 2003; Yager et al., 2007; Recking, 2012; Yager et al., 2012). A wide variety of intrinsic (bed and bank resistance) and extrinsic (sediment production) elements of the channel could combine and limit the sediment supply to the river channel in coarse-grained rivers.

5.4.2. Regional and geomorphological implications

In coarse-bed mountain rivers placed in forested basins, the upstream supply of sediment coming from debris flow and hillslope processes in headwater areas is irregular and sporadic. This fact could contribute to a lack of sediment coming into the channels (Dietrich et al., 1989; Yager et al., 2007; Recking, 2012).

Particularly in rivers draining the Cantabrian Mountain range it has been described a slow degrading trend related to land use changes during the last century (loss of cropping areas, forestation of upland basins) that could be limiting the supply of sediment coming from headwaters to the high-order channels (Fernández et al., 2006; Fernández and Anadón, 2010; Vázquez-Tarrío and Menéndez-Duarte, 2014). This degrading trend is reflected in channel narrowing, loss of secondary anabranches, and vegetation growing in banks and riparian areas (Fernández et al., 2006; Fernández and Fernández, 2008).

In this kind of geomorphological setting, perennial coarse-bed mountain rivers typically are featured by the development of bed armoring and packing (Dietrich et al., 1989; Church and Hassan, 2005). Moreover, structural arrangements and several kinds of bedforms (imbrications, patches, clusters) are not uncommon (Wittenberg and Newson, 2005; Hassan et al., 2008; Venditti et al., 2008). All of these structural and textural features locally enhance hydraulic roughness and bed resistance, increasing the threshold stresses for incipient motion and promoting channel stability and low transport rates (Church et al., 1998; Bathurst, 2007; Hassan et al., 2008; Yager et al., 2012). These facts could strongly constrain the actual bed sediment supply coming from in-channel storages during the more frequent floods (Yager et al., 2007; Recking, 2012).

These bed textures and arrangements are self-formed structures whose development is controlled by sediment availability and the history of previous dominant discharges (Dietrich et al., 1989; Church et al., 1998). Somehow, channels accommodate changes in sediment supply and transport capacity not only by changes in hydraulic geometry but also by changes in bed sediment grain size and texture (Mao et al., 2011). Accordingly, these different patterns of packing arrangements and bed sediment textures introduce a strong heterogeneity in hydraulic resistance, so bed state should be considered as a 'degree of freedom' (Ferguson, 2008) not taken into account in any bedload model.

Therefore, the previous considerations (low sediment availability, unsteady and nonuniform nature of bedload transport, structural arrangements) could explain why the bedload formulas fail when estimating bedload discharge in the studied rivers.

The modern equations, like Wilcock–Crowe or Parker–Klingeman–MacLean, are based on well-defined experimental data taken in coarse-bed channels (field and/or flume data). They introduce complex formulations that take into consideration bed armoring and its breakup during transport episodes. They also take into account the effect of sand content on the sediment mixture (Wilcock, 1993; Wilcock and Crowe, 2003), and they make use of hiding functions in order to catch the dependence of bedload rates on grain size (Parker and Klingeman, 1982; Parker et al., 1982; Parker, 1990, 2008; Wilcock and Crowe, 2003; Wilcock et al., 2009). Thus, this explains why those equations provide better results than the 'classic' ones.

However, the tested 'modern equations' are still not including in its formulation all the features governing bedload transport in the studied channels. This point could be related to the fact that some of the tacit assumptions derived from the 'similarity collapse hypothesis' are only approximated in coarse-bed streams.

In this sense of hiding functions and threshold stresses of a sediment mixture change with grain sorting and sand content (Wilcock and Kenworthy, 2002), none of the studied bedload equations is able to include in a single formulation all the possible settings (de Linares and Belleudy, 2007). Also, other phenomena, apart from relative size effects, could be influencing clast entrainment. In that sense, Kirchner et al. (1990) pointed out the following statement: rather than using single shear stresses for each grain size, it should be more adequate using the distribution of entrainment probability for each grain size, if we aimed to properly consider all the phenomena linked to fluctuations in turbulence and instantaneous shear stresses. This statement made by Kirchner et al. (1990) is not considered in the equations tested in the current paper.

In summary, even the modern equations require some assumptions. In common with the classic equations, the modern formulas are still empirical correlations; and of course, more complex than the classical ones, scaled by the flow and fitted to different bed conditions.

We think that the use of bedload equations for predicting bedload rate needs of equilibrium channels, availability if in-channel sources of sediments and a well-defined alluvial channel geometry and bed texture. Far from this situation, discrepancies between real rates and predicted ones are expectable.

6. Conclusions

In this work, we tested nine bedload discharge equations using bedload transport rates obtained with tracers during six flood episodes that occurred in River Pigüeña and River Coto, two mountain coarse-bed streams from the NW Iberian Peninsula.

With the only exception of the W–C equation in River Pigüeña, none of the assessed equations provided good estimations. All of them overestimated the bedload transport rates; in the case of the classical equations, they overestimated in a range of 2 or even 3 orders of magnitude.

We think that the origin of this overestimation lies in the particular geomorphological conditions of mountain coarse-bed streams in humid conditions belonging to forested basins: in this geomorphological setting, the occurrence of bed armoring and structural arrangements in the bed of river channels, together with a low upstream sediment supply (linked to the forested condition of upland basin areas), determine a supply-limited sediment regime that makes the tested equations not applicable.

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