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Continuous monitoring of bedload discharge in a small, steep sandy channel

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Abstract

This paper reports on bedload flux and texture monitored in a natural, steep, sandy ephemeral channel draining a small gullied sandy watershed, the Barranca de los Pinos (1.32 ha), Spain. Bedload flux was continuously monitored with two independent Reid-type slot samplers; bedload texture was determined from the sediment collected in the samplers. Channel morphology was surveyed with a high spatial resolution with a Terrestrial Laser Scanner.

The monitored instantaneous bedload fluxes are among the highest measured in natural rivers, characterized by high temporal and spatial variability related to the presence of bedforms, shallow bars and sand sheets, and to the reworking of the dry bed

between and at the end of individual flow events. The grain size distribution of the bedload indicates equal mobility; but bedload texture fluctuates, depicting the transport of coarser bar surfaces and of finer-grained anabranch surfaces as well as of the overall bed subsurface.

Key Words: Bedload flux, Sand-bed channel, Ephemeral stream, Steep stream, Reid-type slot sampler, Braiding

1 INTRODUCTION

Bedload transport is a fundamental process shaping stream channels. The interrelationships between water, transport of sediment and bed configuration are complex and the mechanical principles that govern their behaviour are not yet adequately explained (Turowski, 2010). Bedload transport is a challenging area of research due to its high temporal and spatial variability (Gomez, 1984), the interaction of different sizes of bed material (Parker, 2008) and the fact that the transport of coarse sediment itself may change channel geometry (Ashmore, 1991). The knowledge of bedload transport mechanisms is of importance, not only academically to better understand the underlying processes and forms, but also to aid managers and engineers in informed and appropriate decision making concerning river and riverine environments (Lancaster and Grant, 2003).

Bedload measurements have been undertaken under a variety of environmental settings. These include river channels with different grain size distributions (sand to boulder beds), bedforms and bed patchiness, gradients and hydrologic regime (perennial to ephemeral). Measuring bedload transport is expensive, time consuming and also dangerous in some settings, hence measurements of bedload are less common than those

of suspended sediment (Gray et al., 2010). In flumes, bedload has been monitored under controlled conditions with uniform material (e.g., Meyer-Peter and Müller, 1948) and with mixed size sediment (e.g., Iseya and Ikeda, 1987; Recking et al., 2009). In the field bedload transport is difficult to measure due to the complexity of this phenomenon (Haff, 1996) as it entails high spatial and temporal variability, complex grain size distribution and large sizes, high flow velocities and turbid flow. Moreover it is difficult to determine when and how sediment moves on the bed; also, the flow and bed may be disturbed by deployment of bedload samplers (Holmes, 2010). Three types of devices have been used to measure bedload transport in rivers. The first involves the use of portable samplers, such as the Helley-Smith (Helley and Smith, 1971), Arnhem (Schaank, 1937) or Delft-Nile (Van Rijn and Gaweesh, 1992) and portable traps (Bunte et al., 2001). Data obtained with these samplers are limited to a single location for a short time interval, but the samplers are movable between sites. The second includes devices that allow the continuous direct measurement of bedload transport at fixed locations, such as the continuous belt slot system at the East Fork River (Leopold and Emmett, 1977), the channel-wide vortex slot (Milhous, 1973), the ultrasonic sensor system developed in Rio Cordon (D'Agostino and Lenzi, 1999) and the Reid-type recording slot sampler (Reid et al., 1980). The latter has been the most widespread method, having been used successfully in permanent and ephemeral gravel bed rivers worldwide (García et al., 2000; Laronne et al., 2003).

More recently, surrogate monitoring technologies, such as Acoustic Doppler Current Profilers - ADCPs (Gaeuman and Jacobson, 2006a), geophones (Mizuyama et al., 2010; Rickenmann and Fritschi, 2010) and hydrophones (Belleudy et al., 2010) have been developed. These are non-contact devices collecting information indirectly and allowing continuous monitoring of bedload transport under a larger number of

scenarios. However, these technologies are —to some extent— in the experimental phase, and require the collection of physical samples for calibration. Yet, these devices will most likely be those used in the future to collect information on bedload transport (Gray et al., 2010),

Bedload transport in sand-bed systems has been studied mostly in flumes, (Ashmore, 1988; Bagnold, 1966; Einstein, 1950; Engelund, 1966; Engelund and Hansen, 1967), but also with portable bedload samplers in large (Gaweesh and Van Rijn, 1994) and small rivers (Billi, 2011). And more recently by the use of surrogate techniques (Gaeuman and Jacobson, 2006b; Rennie and Villard, 2004). However, datasets on bedload transport in natural ephemeral sand bed channels are relatively rare and largely incomplete (Billi, 2011). This is given to the fact that sand bedded rivers are usually large rivers, with high water discharge and low slopes. In these settings, there are often large bedforms during higher flows, which make the continuous monitoring of bedload impractical (Holmes, 2010).

Among ephemeral rivers, bedload transport has been mostly studied in gravelly beds with higher bedload fluxes compared to their perennial counterparts (Laronne and Reid, 1993); apparently, ephemeral steep sand bed rivers also have high bedload fluxes (Billi, 2011). As the channel bed in ephemerals is unarmoured (Laronne and Reid, 1993), hysteresis in bedload flux is rarely observed (Powell et al., 2003). Single thread ephemeral gravel bed rivers with moderate slopes exhibit a sequence of steeper bars and less steep, finer-grained 'flats' (Powell et al., 2012). In steep sand bed rivers, the bed tends to be flat, lacking bars, ripples and dunes; sheets are often observed in the channel bed (Billi, 2008).

This paper aims to provide continuous bedload observations on a fluvial system that has yet to be reported: a natural steep sand-bedded river with an ephemeral regime.

101 We attempt to understand bedload flux and texture and their relations with hydraulic
 102 parameters. This objective is accomplished by acquisition and analysis of continuous
 103 bedload flux data obtained with two Reid-type slot samplers installed in the stream bed.
 104 We aim to comprehend their spatio-temporal variation, as well as to which extent
 105 bedload flux varies with shear stress.

106

107 **2 STUDY AREA**

108 The Barranca de los Pinos is located in the Northern piedmont of the Guadarrama
 109 Mountains, Segovia Province in Central Spain. The underlying topography consists of a
 110 series of mesas and cuevas formed by Upper Cretaceous sediments and underlain by a
 111 crystalline basement of gneisses (Fig. 1). The plateaus are topped by a caprock of
 112 limestone and dolostone while the side slopes are clayey and gravelly sands that have
 113 been deeply dissected by gullies. The mesas and cuevas are covered by native forest
 114 (holm oak and junipers) and are grazed by sheep at certain periods of the year. The
 115 climate is Mediterranean with cool summers (Csb) according to Köppen classification
 116 (CNIG, 2004). It is characterized by a moderate average annual precipitation (680 mm)
 117 and temperature (11.4°C).

118 The Barranca de los Pinos is typical of the gullied catchments of the studied area
 119 in terms of size, lithology and gradient of hillslopes, and channel. It has been chosen to
 120 study different active geomorphic processes: gravitational processes in high gradient
 121 slopes, water erosion on low gradient slopes and sediment transport including bedload
 122 in the channel (Lucía et al., 2011). The catchment area (1.32 ha) is to a large extent
 123 gullied (90.4%), with high gradient slopes (29.9% of the gullied area has slopes steeper
 124 than 30°), narrow interfluvies and a high drainage density (0.041 m m⁻²). The
 125 longitudinal slope of the channel is 0.066 and its width at the monitoring site is 1.24 m,

126 varying in the range 1 - 1.5 m. The gullied reach has friable, vertical sandy walls, but at
 127 the sampling site they are stable (Fig. 2); the slopes of the right and left banks are 29.4°
 128 and 78.8°, respectively and their height is approximately 70 cm. The channel bed lacks
 129 topography or undulations with a maximum 2 cm relief. It is of low sinuosity, 1.08,
 130 classified as straight to slightly sinuous (Leopold et al., 1964). The bed is formed by
 131 coarse, poorly sorted and positively skewed sand ($D_{50} = 0.555$ mm; $D_{84} = 0.995$ mm;
 132 $D_{90} = 1.42$ mm). Most (93.2%) of the bed material is sand sized, ranging from 0.062 to
 133 2 mm . There are very small proportions of silts and clays (2.9%) and of gravel (3.9%),
 134 the latter being subrounded to angular quartzite lag deposits from the sandy facies or
 135 very angular carbonate rock fragments originally derived from the caprock and the
 136 associated colluvium.

137 Steep, sandy channels are uncommon in Nature, as finer grained rivers usually
 138 have lower gradients (Leopold et al., 1964). The studied channel is a rare combination
 139 of a sand bed and steep longitudinal slope. It exists here because the gullies are
 140 presently eroding fine-grained Upper Cretaceous sediments, deposited by large braided
 141 and meandering rivers in an estuary mainly by fluvial but also by tidal activity (Alonso,
 142 1981), conditions which are indicative of low gradient channels. The gullied character
 143 of the catchment provides an unlimited sediment supply to the channel. The study area
 144 has been described in detail elsewhere (Lucía et al., 2011).

145

146 **3 METHODS**

147 **3.1 Water stage**

148 Water stage was measured at the study site by a vented pressure transducer located in
 149 one of the bedload slot samplers. Water density was assumed constant at 1043 kg m⁻³.

150 **3.2 Bed topography and texture**

151 The topography of the channel bed was acquired from a point cloud data obtained with
152 a Terrestrial Laser Scanner (TLS), which is based on Light Detection and Ranging
153 (LiDAR) technology. The TLS is non-intrusive and has high precision; the instrument
154 used (Leica Scan Station 2) measures up to 50,000 points per second with a 2 mm
155 precision at a scanning distance <120 m. Scanning was undertaken at least from two
156 different locations to avoid shadowed areas (Buckley et al., 2008). The scanned channel
157 reach is 12 m long (ten times the channel width) and it is located immediately upstream
158 of the bedload sampler. The slope of the channel banks formed an asymmetric
159 trapezoid, from which the hydraulic radius was calculated.

160 To determine the bulk Grain Size Distribution (GSD) of the bed-material, an
161 11.6 kg sample was scraped from the upper 1–2 cm of a 2 m^2 area of the channel bed.
162 Only one area was sampled given the uniformity along the channel; nonetheless, this
163 area is longer than the channel width. The sample was dried and sieved at 1 intervals
164 and lower-truncated at 0.062 mm (sand-silt split). Grain size descriptions were
165 calculated using Gradistat (Blott and Pye, 2001).

166 **3.3 Bedload**

167 Bedload discharge was automatically and continuously monitored by two independent,
168 cross-sectionally aligned Reid-type (formerly termed Birkbeck) bedload slot samplers
169 (Reid et al., 1980). The cumulative mass of sediment entering each sampler is
170 monitored by a vented pressure transducer connected to a pneumatic pillow filled with
171 water, upon which an internal box is located. The hydrostatic pressure of the water
172 column is monitored by a separate vented pressure transducer located between the outer
173 and inner boxes of the right sampler. For a given time period, the pressure difference
174 between the two sensors is due to the addition of mass of bedload entering the sampler

(Laronne et al., 2003). Data from all the vented pressure transducers are read every 10 s and the average of three readings is logged every 30 s.

The volume of each inner box is 0.225 m³ and was sized based on prior sediment yield assessment so that the box would not overflow during a typically frequent, low magnitude event (Lucía et al., 2011). Slot width is variable, the maximum (160 mm) representing 26% of the channel width. Ideally the slot width should be ten times larger than the size of the sediment to be sampled, and a compromise is required between the sampled sediment diameter, the representative width and the average sampling duration required for samplers to fill. During the sampling period the slot was set at 5 or 10 cm, much larger than the bed material grain size (50-100 times the D₈₄). The length of the slot was based on the saltation length for sand calculated as follows (Van Rijn, 1984):

$$\lambda_b/D = 3D^{*0.6} T^{0.9}$$

where λ_b is the saltation length with an accuracy of 50%, D is the particle diameter, D* the dimensionless particle parameter defined as:

$$D^* = D_{50}[(\rho_s/\rho_w)g/v^2]^{1/3}$$

where ν is the kinematic viscosity, ρ_s and ρ_w are the mass densities of sediment and water, g is gravity and T the transport stage parameter, which is defined as:

$$T = [(u^*)^2 - (u_{cr}^*)^2] / (u_{cr}^*)^2$$

where u^* is the bed shear velocity equal to $g^{0.5}/C'$, C' is the Chézy coefficient related to the grains, and u_{cr}^* is the critical shear velocity according to Shields (equal to $[\theta_{cr}^*((\rho_s/\rho_w)-1)g D_{50}]^{0.5}$). Applying this equation and using the D₅₀, 0.04 as the Manning parameter (Arcement and Schneider, 1989) and a maximum water depth of 30 cm, the saltation length was estimated to be 36 cm (±50%); therefore, the maximum length was predicted to be 54 cm. Indeed, the slot length at 65 cm is sufficient for the predicted transport conditions, thus having a 20% safety factor.

200   The sampler has a lateral window allowing observation of sediment
201 stratification, hence enabling the collection of facies-based sediment samples. In all but
202 few cases, sample weight was 100 times larger than the weight of the largest particle as
203 recommended by Church et al. (1987). Bedload samples collected before April 2010
204 were dried and sieved with 1 sieves and those collected after this date were sieved
205 using 0.5 sieves because it became evident that finer textural detail was required. The
206 smallest sieve size used was 0.062 mm, essentially the lowest truncation. As these
207 bedload samplers are of the recording type, the time during which a given layer of
208 bedload sediment was deposited can be determined. This allowed correlating bedload
209 texture with the channel-average shear stress, typifying the hydraulic conditions existing
210 when the sediment was transported (Powell et al., 2001).

211   All vented pressure transducers (Druck PTX-1830) were pre-calibrated. The
212 sensor measuring hydrostatic water pressure has a sensitivity of 0.06% (according to the
213 manufacturer). The sensitivity of this kind of weighting device comprising the pressure
214 transducer and the pillow is estimated to be 0.3 kg given its size, (Laronne et al., 2003).
215 However, sensitivity was also tested during calibration of this sampler. The calibration
216 was undertaken with metal pieces of known weight larger than 9 kg, as well as smaller
217 objects (0.25, 0.5, 1, 3 and 5 kg) while the sampler had a weight equivalent to (i) being
218 filled only with water, (ii) when half-filled with sediment and (iii) when almost full with
219 dry sediment. In all cases there was a significant linear correlation between the pressure
220 registered and the weight introduced to the sampler. Regression lines were compared
221 one to one with the regression line obtained with weights larger than 9 kg. The
222 comparison was made using an ANOVA test, analyzing both the slope and the intercept.
223 In this analysis, the slope is more relevant because it is the extent to which pressure
224 changes with weight in the sampler. Obtained p-values show that there are no

statistically significant differences among the slopes or the intercept at the 90% or higher confidence level to an accuracy of 0.25 kg. This value was used to analyse the data: increments of weight smaller than 0.25 kg were excluded in calculation of bedload flux.

The temporal stability of this type of pressure transducer has been shown to be quite good (Alexandrov et al., 2009). However, the properties of the pillow, which is made of neoprene, may not be similarly stable. Therefore, the weighting device was calibrated twice (in April of 2010 and February 2011), with results showing little change in the slope of the regression lines: 3.08% in the right sampler and 10.1% in the left sampler. Hence, for the events prior to August 2010 (the date in between the two calibrations) the regression line obtained during the first calibration was used to predict transported mass and that obtained with the second calibration was used thereafter.

3.4 Bedload data quality control

The reproducibility and quality of the bedload database is validated by the following procedure:

1. At the onset of some very small flow events calculated bedload flux rates were excluded due to two known errors: For a correct flux calculation the sampler has to be filled with water, which may take a few minutes (one to five) during small events. In other instances, the collected data were unrealistically high due to the effect of bed over-steepening that was produced by the process of cleaning of the upstream section of the sampler, after the previous event was recorded.
2. Slot sampling has been demonstrated in the lab (Poreh and Sagiv, 1970) and under field conditions (Habersack et al., 2001) to be 100% efficient while sampling sand to gravel bedload as long as they are not nearly full. When the volume of sediment in the sampler approaches about 80% of its capacity, the

250 efficiency decreases due to internal vortices that can remove some of the
 251 sediment from the samplers (Habersack et al., 2001). Data collected under these
 252 conditions were removed due to sampling inefficiency.

253 3. During the latter stages of some hydrograph recessions, water depth over the
 254 sampler decreased much slower than expected, most likely due to sand
 255 deposition over the sampler or mud deposition within the sampler on the
 256 pressure transducer. This effect was corrected by adjusting hydrograph recession
 257 up to the inflection point to an exponential equation. From the inflection point
 258 onwards, bedload data were correlated with depth recalculated by the
 259 exponential equation.

260

261 **4 RESULTS**

262 **4.1 Hydraulics**

263 The Barranca de los Pinos is a truly ephemeral channel. Water was present in the
 264 channel during 1.98% of the monitored time, merely 11 of 556 days. Twenty four flow
 265 events were registered during 18 months, June 2009 – January 2011 (Table 1). All the
 266 events were generated by rainfall with a return period smaller than two years. Three
 267 events (9, 22 and 23) did not register accurately as they were small and short, with
 268 difficulties that occur at the beginning of some events (see section 3.5) affecting the
 269 entire event dataset; these were excluded.

270 Maximum registered water depth was 15.5 cm (averaged every 30 sec). Median
 271 water depth was 2.6 cm, and the first and third quartiles were 1.0 cm and 4.5 cm
 272 respectively. The longitudinal slope of the channel is considerable (0.066), so despite
 273 the shallow water depth, shear stress was quite high, with a maximum instantaneous (30
 274 s) value of 10.7 N m^{-2} . Channel average shear stress was calculated as (Du Boys, 1879):

$$\tau = \rho_w d S g$$

where τ is shear stress (N m^{-2}), d is water depth (m) and S is bed slope (nondimensional).

4.2 Bedload flux

During most of the events bedload transport was initiated soon after water appeared in the channel. On average the onset occurred within 4.6 and 6.2 min in the right and left-hand samplers, excluding few events in which the hydrograph rise was exceptionally slow (inclusion of these events increases the respective average gaps to 13.1 and 26.6 min).

One of the limitations of the Reid bedload sampler is its finite volume. However, given the small size of the catchment and the brevity of some of the bedload-generating flow events, there were nine events when both samplers did not fill entirely, and one additional event during which only one of the samplers did not entirely fill. Bedload flux was monitored at peak flow in six of 12 events when the samplers had filled.

When samplers were full the maximum cumulative mass varied between the samplers (Table 1). This value is the integration of bedload fluxes for the sampling period before sampler efficiency decreases. As bedload fluxes lower than the sensitivity (0.25 kg in a given time interval) are excluded, the samplers may contain more sediment than that calculated as the total cumulative mass.

Measured bedload fluxes were high for both samplers; the highest 30 s recorded values were 25.4 and 19.5 $\text{kg s}^{-1}\text{m}^{-1}$ in the left and right sampler, respectively. During the monitoring period, 2375 values of 1-minute averaged bedload flux were obtained; their median, first quartile and third quartile were 0.33, 0.16 and 0.70 $\text{kg s}^{-1}\text{m}^{-1}$. In a general sense, the relation between bedload flux and water depth may be simple or complex (Cohen et al., 2010). In the analysed database (1-min averaged), the relation

300 between the entire bedload flux vs shear stress is very scattered (Fig. 3a). However,
 301 lower scatter characterizes some events (Fig. 3b).

302 **4.2.1 Temporal variation**

303 Given the large scatter in bedload flux, the variability of bedload transport rates is
 304 examined by evaluating two types of temporal variability: hysteresis (variations in
 305 bedload flux on rising vs falling hydrograph limbs) and waves (periodic fluctuations in
 306 bedload flux unrelated to changes in flow stage).

307 The rate of change of water stage is considerably more rapid during the rising
 308 limb than during the recession in many of the events. Hence most of the bedload flux
 309 data were obtained from recessions by virtue of this portion of the hydrograph lasting
 310 longer (Fig. 4). Observed instances of hysteresis in the variation of bedload flux with
 311 water depth are without exception clockwise, with higher rates of transport occurring
 312 during rising stage than during flow recession. This was documented in eight and nine
 313 events among twenty in the left and right sampler, respectively. Bedload hysteresis was
 314 observed mostly in summer and autumn. The proportion of events with hysteresis is
 315 highest (83%) in the summer, and lowest (20%) in spring, followed by winter (30%)
 316 and somewhat more (42%) in autumn (Fig. 5).

317 In some of the monitored flow events, or parts thereof, bedload flux
 318 corresponded well to water depth (Fig. 6a). At other times, large oscillations of bedload
 319 flux (waves) occur both, during steady flow (Fig. 6b) and unsteady flow (Fig. 6c).
 320 Oscillations were documented in eight and nine events in the left and right samplers
 321 respectively, indicating the frequency of waves in bedload response; the presence of
 322 waves is independent of water depth. Notably, hysteresis and waves do not necessary
 323 occur simultaneously. Wave occurrence varied seasonally less than did hysteresis.
 324 Waves occurred in 40% of the spring events and in 67% of the summer events (Fig. 5).

325 **4.2.2 Spatial variation**

326 Spatial variation of bedload flux was described based on the evaluation of
 327 registered bedload flux differences between samplers and, separately, their temporal
 328 responses. Considerable spatial variation (differences in bedload rates of more than the
 329 50% or more than 5 minutes of interval in bedload flux registration) occurred in 11 of
 330 20 events. In nine of the 11 events, bedload flux occurred later in the left sampler
 331 compared to the right sampler. Bedload entrainment was recorded in the left sampler
 332 when water depth attained a minimal threshold depth in the range 17-35 mm. The
 333 largest spatial differences in bedload flux occurred in shallow, bedload-transporting
 334 flows during hydrograph rise (Fig. 7).

335 **4.3 Bedload texture**

336 Bedload collected in the Reid-type samplers showed an alternation of coarser and finer-
 337 grained sedimentary layers. Bedload texture was analysed from 276 facies-based
 338 bedload samples. Correlating the thickness of the various facies with their cumulative
 339 weight allowed inferring when the sample was collected (Laronne et al., 2003). The
 340 GSD of the samples were averaged and compared to the corresponding shear stress in 5
 341 N m^{-2} bins. Interestingly, the GSD of bedload is unrelated to shear stress (Fig. 8)
 342 indicating that selective transport cannot be deduced from these data. The range in D_{50}
 343 variation is smaller than that for D_{90} , but the relative variability is similar (Fig. 8), as
 344 expected given their respective sizes (Whitaker and Potts, 2007).

345 **4.4 Morphotexture of the channel bed**

346 The explanation of (1) the alternation of GSD facies within the sampler despite non-
 347 selective bedload transport, as well as (2) the spatiotemporal variation of bedload flux
 348 while flow depth remained essentially constant, appears to depend on the character of
 349 channel bed morphology. Comparison of the median water depth with the relief of what

at first appeared to be a simple flat channel with minute topographic differences, in fact shows that both are of similar magnitude. A zoom into the ephemeral channel bed after the occurrence of a bedload-generating flow event reveals that the bed is comprised of bedforms with an apparent braided pattern (Fig. 9a). To describe the characteristics of the channel bed in detail, a topographic survey was carried out with the Terrestrial Laser Scanner (TLS). A 10 m channel reach of a tributary gully of the Barranca de los Pinos was selected for this survey because the topographic characteristics are essentially identical in both channels, and because the Barranca bed was disturbed by animal trampling which destroyed its micro-topography. The scanned tributary joins with the main stem immediately downstream of the Barranca monitoring station.

The DEM obtained with the high resolution (1 mm) topographic survey demonstrates that the channel has a well-defined braided pattern (Fig. 9c) with complex bars on which chutes are developed, having an average length, width and height of 91, 21 and 1.2 cm respectively. The average braiding index, defined as the number of anabranches (arrows in Fig. 9d) per cross section (Egozi and Ashmore, 2008), is 3.5 (Fig. 9d).

Considering the presence of these bedforms, a new sampling strategy was undertaken to better characterize channel texture (Fig. 9b). Bar and anabranch surfaces and subsurface were separately sampled, as was the general subsurface (Fig. 10a). The sampling of the surface was undertaken by carefully scraping one-grain layer of surface sediment. The subsurface was characterized by a bulk sample representing 1-2 cm of the subsurface sediment. A large (3.4-fold) difference in grain size occurs between the D_{50} of the surface of anabranches (0.39 mm) and that of the bars (1.30 mm). That bar surfaces are coarser-grained than the subsurface indicates that the bar surface is affected by a phenomenon of segregation which is absent in the anabranches. The median of the

375 bar subsurface tail is 20% finer than the respective bar head, revealing the existence of a
376 bar-scale sorting process.

377 Comparing the GSD of the different parts of the channel with the samples of the
378 bedload retained in the samplers shows that bedload texture for many of the samples
379 was both coarser than that of the anabranch subsurface and finer than that of the bar
380 surface (Fig. 10b). Nearly half (44.5%) of all bedload samples were finer-grained and
381 the rest, a slightly larger fraction, coarser grained than the anabranches. Only two
382 bedload samples had a larger D_{90} than the respective centile of the bar surface. The
383 frequency of movement of the different sizes of bedload was analysed considering the
384 individual sampling duration and the total event sampling duration of each of the right
385 (RS) and left (LS) samplers (Fig. 10b). The GSD of the bedload collected in both
386 samplers is almost identical to the GSD of the average of the channel to 2 cm depth,
387 demonstrating that equal mobility characterized the entire duration of bedload
388 monitoring. However, since the analyzed bedload samples show an alternation of
389 coarser and finer-grained layers, and bedload transport was not selective with reference
390 to increasing shear stress (Fig. 8), it is suggested that the observed variations in GSD of
391 the bedload are related to bedform movement.

392

393 **5 ANALYSIS**

394 **5.1 Prediction of bedload flux**

395 In an attempt to determine the applicability of bedload equations to small, steep sand
396 bed channels, monitored bedload flux data were compared to selected bedload equations
397 (Fig. 11). Ten minute averaged data were used to diminish temporal variability inherent
398 in bedload transport (Ergenzinger et al., 1994). The ratio between calculated and
399 observed values ranged as much as three orders of magnitude.

400 The Smart and Jaeggi (1983) equation, established for flows on steep slopes and
 401 nearly uniform sediments including sand, was first compared with our data. The
 402 formula considerably overestimates with a median calculated/measured ratio equal to 16
 403 (Fig. 11). The Smart and Jaeggi formula was established for straight channels with
 404 minimal bedforms, whereas in our study bedload was measured in braided channels,
 405 where form resistance is far from negligible.

406 Despite being developed for lower slopes and coarser sediment, the well-known
 407 Meyer-Peter and Müller or MPM (1948) formula was also examined as it has become
 408 a standard for estimating bedload under a variety of settings. Bedload was
 409 overestimated with a median ratio of 10 (Fig. 11). An improved fit, the range of
 410 discrepancy decreased to about two orders of magnitude (corresponding to a median
 411 ratio of one) was obtained by including the roughness correction $n'/n=0.4$ (where n' is
 412 the grain roughness and n is the total roughness); however flow velocity data were
 413 unavailable to assess the appropriateness of this value.

414 The Ashmore (1988) equation was developed from data obtained in flume
 415 experiments with conditions similar to the ones present in the Barranca de los Pinos
 416 channel (sand and small gravel $D_{90}=4$ mm, though with a gentler slope 0.01-0.015). It
 417 was developed as a model for braided gravel bed rivers. The results show an
 418 overestimation, with a median calculated/measured ratio of 3.6 (Fig. 11). This is not as
 419 large as other ratios, well within the -0.1 to 10 range recently used for similar
 420 comparison of bedload equations (Recking, 2010) in consideration of the uncertainties
 421 of the empirical equations and of the queries associated with bedload measurements (see
 422 hereafter). This equation implicitly takes into account form resistance associated with
 423 the braided pattern, with no requirement for an *a priori* hydraulic correction as with
 424 MPM. However, it was derived for the mean bed shear stress (calculated from the cross-

sectionally averaged depth and width), whereas in this study, local depth (that over the right slot sampler) was used.

In summary, such equations were expected at best to predict the median bedload flux, though with admittedly large confidence intervals. Certainly none of these and other tested bedload formulae can be expected to reproduce the large variations about median (or mean) bedload fluxes, fluctuations which are inherent to bedload transport in multithread channels.

5.2 Fluctuations in bedload flux

One of several reasons for variability in bedload transport is the fluctuating nature of boundary conditions at a given location: slope (S), grain diameter (D) and flow depth (approximately equal to the hydraulic radius (R) for shallow flows); these are the building blocks of the Shields parameter or non-dimensional shear stress (*):

$$* = R S / [D((\rho_w/\rho_s)-1)]$$

where R is hydraulic radius. These parameters varied in time and space in this study as follows:

150% for sediment diameter when considering maximum and minimum D_{50} measured in the different parts of the channel (Fig. 10a);

$\pm 0.8\%$ of the average slope – the maximum fluctuation observed in some flume experiments with high longitudinal slope (9%) (Recking et al., 2009);

water depth minus a range from 0 to 3 cm; 3 cm is the maximum bar height in the cross section (Fig. 9d).

It is relevant to note that the calculated variations in these parameters do not completely explain the large variability of observed bedload flux. Indeed, bedload fluctuations are also linked to variation in the supply of sediment and occur in rivers under steady flow

(Ashmore, 1988; Cudden and Hoey, 2003; Ergenzinger et al., 1992; Gomez, 1983; Gomez et al., 1989; Recking et al., 2009; Turowski, 2010).

To determine whether such fluctuations occur and also their nature, frequencies of the temporal variation of bedload flux and water depth were analysed using a Fourier transformation (Recking et al., 2009). For this analysis, only data from the right bedload sampler were used, because the vented pressure transmitter recording water depth is located in the right slot sampler, and there are considerable variations in water depth across the channel at shallow flows. Most of the events had a short duration, which prevented undertaking a thorough Fourier analysis for all the events. Four events (7, 8, 20 and 23 – Fig. 12) had a duration longer than 100 min, considered to be sufficiently long to permit a time series analysis. These were sampled at 1 min interval. The four events represent a range of flow characteristics while bedload flux remained within a similar range of values. Although a clear peak of frequencies is not observed, all the bedload flux signals have an identical spectral signature (Fig. 13b) despite the different frequency spectrum of flow depth (Fig 13a). This suggests that fluctuations are in part controlled by internal mechanisms such as bedform movement. While there were no clear peaks in the bedload signal there was a progressive evolution covering all frequencies, indicating that the phenomenon responsible for fluctuations is not discrete, but continuous; e.g., bedload sheet movement or the braiding pattern, which incessantly changes over time.

469

470 **6 DISCUSSION**

The obtained data allows characterizing bedload flux and GSD and its relation with the shear stress in this environment, revealing a complex system with several particularities.

473 **6.1 Bedload flux and hydraulics**

474 The Barranca de los Pinos is distinctly ephemeral, with water and sediment movement
 475 occurring during only about 2% of the study period, similar to many other ephemeral
 476 streams (Reid et al., 1998). During this period, the mean water depth in the channel was
 477 16 mm, ranging from 1 to 155 mm. Despite the shallow flow, bedload fluxes were high;
 478 the 1st and 3rd quartiles were 0.33 and 0.70 kg s⁻¹m⁻¹ but maxima of more than 20 kg s⁻¹
 479 m⁻¹ were registered. These bedload fluxes are higher than fluxes continuously
 480 monitored in perennial gravel bed rivers in different environments, ranging between
 481 0.001 and 1 kg s⁻¹m⁻¹ with maxima rarely higher than 1 kg s⁻¹m⁻¹ (García et al., 2000;
 482 Habersack et al., 2001; Laronne and Reid, 1993; Mao et al., 2010; Milhous, 1973;
 483 Rickenmann and McArdell, 2007; Vericat and Batalla, 2010). They are also higher than
 484 the few measured rates in small flow events in an ephemeral sandy river having a steep
 485 longitudinal slope, the Gereb Oda (Billi, 2011) where measured bedload ranged from
 486 0.01 to 1 kg s⁻¹m⁻¹.

487 Bedload fluxes obtained in the present study are comparable to those measured
 488 in sandy gravel bed rivers draining active volcanic terrain such as Mt. Pinatubo after its
 489 eruption (Hayes et al., 2002) with rates from 0.1 to 2.2 kg s⁻¹m⁻¹ and to upland
 490 ephemeral, gravel bed rivers in the Israeli desert: Nahal Eshtemoa (Reid et al., 1998);
 491 Nahal Yatir (Reid et al., 1996) and Nahal Rahaf and Qanna'im (Cohen and Laronne,
 492 2005), where respective transport rates of 0.1 to 2.2, 0.01 to 8, 0.1 to 37 and 0.1 to 15
 493 kg s⁻¹m⁻¹ have been measured, in four ephemeral gravel bed rivers with an identical
 494 method and a 1-min averaging duration of bedload flux. In fact, bedload fluxes in the
 495 Barranca were in a similar range and produced by the same magnitude of shear stresses
 496 as in these ephemerals. While channel types are distinct there are similarities: they have
 497 a segregated coarser bar surface, almost twice the median size of the subsurface (1.71
 498 and 1.98 times coarser in the Nahal Yatir and in the Barranca respectively). However, in

499 the Barranca de los Pinos the slope is steeper whereas sedimentary grain size and water
500 depth are at least one order of magnitude smaller. The reasoning for the high Barranca
501 bedload fluxes is thought to be the ephemeral character of the channel (Laronne and
502 Reid, 1993), the fine texture of the channel bed, the steep longitudinal slope and the
503 high sediment supply (in the sense of Dietrich et al., 1989).

504 Ephemeral rivers continuously monitored using Reid bedload samplers have
505 been shown to have a high correlation between channel average bedload flux and cross-
506 sectional averaged shear stress. Where cross-sectional variations do occur, they are
507 ascribed to variation in local shear stress (Powell et al., 1999). The dependence of total
508 bedload yield on average shear stress is also strong in miniature braided sandy channels
509 formed in flumes (Ashmore, 1988). However, in most of these relations a substantial
510 scatter was evident, as is in the channel of the Barranca de los Pinos. In this site, the
511 scatter is explained as a consequence of two types of temporal variation (hysteresis and
512 sediment waves) and spatial variability.

513 Despite substantial spatial and temporal scatter, measured bedload flux data
514 were compared to a set of standard bedload equations to evaluate the ability of these to
515 predict rates of bedload flux for braided sandy streams. Standard bedload equations tend
516 to underestimate bedload sediment yield when they are used with width- averaged input
517 data because they are non-linear (Gomez and Church, 1989; Ferguson, 2003; Paola,
518 1996); this is particularly true for braided rivers with highly irregular sections (Bertoldi
519 et al., 2009; Nicholas, 2000). The contrary (overestimation) was observed here when
520 estimates from the Meyer-Peter and Müller and Smart and Jaeggi equations were
521 compared against the Barranca de los Pinos database. This can be explained by two
522 reasons: first, calculations were not made with the width averaged data, but with a local
523 shear stress computed from the depth measured at the right slot sampler. Second, the

524 computed shear stress was not corrected for form-induced resistance, which was likely
 525 higher than the grain shear stress. The empirical Ashmore (1988) equation developed in
 526 a flume for gravel bedded braided rivers predicts better the bedload flux response, even
 527 though it is to be applied to channel average values rather than to local bedload flux.

528 **6.2 Morphotexture of the channel**

529 The Barranca bed topography, as well as the temporal and spatial variability in bedload
 530 flux and its texture, point to the existence and importance of bedforms. Bedforms are a
 531 result of the interactions between coarse and fine fractions during bedload transport of
 532 poorly sorted bed material (Dietrich et al., 1989), as observed in experiments at constant
 533 water discharge in flumes (Ashmore, 1988; Iseya and Ikeda, 1987; Nelson et al., 2009;
 534 Recking et al., 2009) and in sandy natural rivers (Whiting et al., 1988).

535 The observed bedforms in the channel of the Barranca de los Pinos could be bars
 536 or sand sheets. Given their average dimensions: 91 cm long and 1.2 cm thick, they are
 537 to be considered bars since their size is larger than the dimensions given for bedload
 538 sheets – a length of 100 to 600 grains and one or two grains thick (Whiting et al., 1988),
 539 which, scaled to the studied channel, would be equivalent to 0.6 m long bedforms with a
 540 thickness of 2 mm. The bedforms are also more extensive than bedload sheets observed
 541 in the Gereb Oda, (Billi, 2011). The thickness of the bars in the Barranca is almost half
 542 of the median water depth, similar to the height of bedforms described as bars that were
 543 present in flume runs of braiding using sand (Ashmore, 1982). We have observed the
 544 activity of these features during bedload generating events: they move and reform
 545 similar to bars observed in flumes, but we have insufficient observations to state more.

546 From the available information, we deduce that the Barranca has two bedforms :
 547 bars (based on the topography and bed material texture) as well as somewhat smaller
 548 bedload sheets (based on oscillations/waves of bedload flux with time and the texture of

549 the bedload) moving over more stable bars in a braided pattern. This pattern has been
550 observed in flume experiments (Ashmore, 1988; Hoey and Sutherland, 1991) and in
551 gravel bed rivers (Church and Jones, 1982; Rice et al., 2009). Indeed, bars are formed
552 by the accumulation of successive bedload sheets (Rice et al., 2009). The topographic
553 signature of the sheets is not distinguished in the field nor in the TLS-based DEM;
554 nonetheless, they do contribute to the roughness detected on the bars (Fig. 9c). It is
555 apparent that bars are reshaped by the flow in the anabranches during the recessions as
556 we have observed in the few instances while present during recession and as suggested
557 elsewhere (Billi, 2008).

558   The bar surfaces, which are coarser-grained relative to the subsurface, indicate
559 the occurrence of segregation, a phenomenon observed in some gravel-bed channels and
560 explained by *en masse* deposition, particularly of the coarser sedimentary particles
561 (Duncan and Laronne, 1998) or else by the winnowing of fines (Leopold, 1994). The
562 equal mobility and the non-size selective transport (Batalla and Martin-Vide, 2001) of
563 Barranca sandy bedload indicates that the segregated surface is unstable (in the sense of
564 Gomez, 1984). Indeed the one-particle diameter surface layer of the bars is not well
565 packed, having no observed interlock. The coarser surface has been described as
566 resulting from bedload sheet transport (Recking et al., 2009). The latter is assumed to
567 result from a kinetic sieving process, being a very efficient sorting mechanism which
568 occurs in a moving layer, where the fine fraction is driven downward into the sediment
569 deposit and thereby produces a coarse bed surface (Frey and Church, 2012).

570   Compared to the segregated, sandy Barranca bars, those in ephemeral gravel bed
571 rivers have been shown to be to a large extent unsegregated (Laronne et al., 1994). The
572 miniature anabranches which are unsegregated, typical of other ephemeral systems
573 (Hassan et al., 2006; Laronne et al., 1994), have been explained to form by high

574 sediment yields and rapid recessions that minimize sediment winnowing. The processes
575 occurring on the channel bed during bedload transport appear to be similar to those
576 described in ephemeral gravel bed rivers. That the subsurface in the bar tail is finer-
577 grained than in the bar head reveals that bar-scale sorting processes also occur, however
578 apparently not as efficiently as in gravel-bed rivers (Rice and Church, 2010). The lesser
579 textural gradient may owe its character to the finer overall texture and the better sorting
580 in the Barranca de los Pinos.

581 **6.3 Interaction between morphotexture and bedload flux variability**

582 Explanations for clockwise hysteresis in bedload transport are manifold: long lasting or
583 very intense flow exhausting the stored sediment, limited available sediment supply
584 (Humphries et al., 2012; Williams, 1989), sediment delivery from the channel bed and
585 banks or areas adjacent to the channel rather than from upstream sources and lack of
586 channel bed armouring (Hassan et al., 2005). However the Barranca has virtually
587 unlimited sediment supply, so limitations on sediment availability cannot explain the
588 hysteretic response. One process that may generate the clockwise hysteresis is the
589 destruction of the low relief of the Barranca bars between flow events. If so, bed
590 roughness will be lower and water velocity higher at the onset of the following event,
591 which may explain the clockwise hysteretic behaviour of bedload flux. As bedload
592 transport commences, bedforms are reformed to the braided pattern, increasing
593 roughness and decreasing bedload rates.

594 There are several mechanisms through which bedforms may be disturbed
595 between events. Observed animal trampling between flow events did destroy bedforms
596 above the site. Trampling increases roughness by giving rise to hoof-generated
597 indentations. Increased roughness due to trampling would thus result in lower bedload
598 fluxes during the rising limb, so trampling cannot explain the observed results. A

second relevant mechanism is the loss of the minuscule cohesion of the sandy surface during a dry spell between flow events, when the subdued bedforms are blurred by small gravitational movements along their borders, or by aeolian activity, in part removing sediment from the bars and filling the minute anabranches (Good and Bryant, 1985). This could also explain why the braided pattern of the channel was unnoticed before initiating the monitoring of water and sediment in the Barranca. Our data stands to support the second above mentioned mechanism, because hysteresis is only present in events occurring at least eight days after a preceding event. This may explain why the proportion of hysteretic events is higher in summer than in other periods of the year, since in this season rain events are more sporadic and the channel is dryer, meaning less cohesion in the sandy bed surface. Relevantly, at the onset of some events, the GSD was similar in both samplers, which may point to the existence of as yet undeveloped bedforms.

The clockwise hysteresis in the Barranca cannot occur only due to a reduction of the roughness during dry periods since this has been documented in natural rivers (Gaeuman, 2010) and also under controlled flume conditions, with unlimited sediment supply and nonuniform sediment. In the latter, the explanation has been the reorganization of the bed surface, reducing the mobility of the finer sediments, thereby decreasing bedload flux during the falling limb of the hydrographs (Mao, 2012). Therefore, reorganization of the bed surface at the studied site may also reduce the mobility of the finer sand, since the bar surfaces are coarser.

The observed spatial variability in bedload flux, when bedload was registered in the left sampler only when water depth passed a threshold (see example Fig. 7), may occur due to the presence of a bar, the bifurcation of which blocks water from flowing to the left side at shallow depths. When water depth exceeded bar height, bedload was

624 registered over both samplers - over the entire 'braidplain' - reducing lateral differences
 625 in bedload flux. This phenomenon is not always observed, possibly because the bar was
 626 not developed in that position or because it was blurred by inter-event drying or
 627 trampling, as explained above.

628

629

630 **7 CONCLUSIONS**

631 Local but continuous bedload flux data obtained in the Barranca de los Pinos are the
 632 first available for natural sand-bedded channels. Their availability allow a glimpse into
 633 the understanding of bedload transport in steep sandy channels, making headway in the
 634 identification of the sources and causes of temporal and spatial variability.

635 1. Recorded bedload fluxes are among the highest measured to date comparable to
 636 those registered in upland ephemeral gravel bed rivers or rivers draining active
 637 volcanic landscapes, produced by high longitudinal slopes with fine-grained
 638 channel bed material, indicating high supply of sediment.

639 2. The local bedload flux *vs* local shear stress database is characterized by a very
 640 large scatter. Comparisons with bedload equations, even if developed for similar,
 641 though channel average conditions, will predict a relationship that can differ as
 642 much as an order of magnitude from measured values.

643 3. The scatter in bedload flux is produced by the existence of often unrecognized
 644 miniature bedforms: bedload sheets moving over a subdued braided pattern,
 645 thereby producing temporal and spatial variability in bedload flux.

646 4. These bedforms move and evolve during bedload-generating flow events, leading
 647 to sediment waves interpreted as miniature bars with overriding bedload sheets.
 648 The presence and emergence of very small central bars is the mechanism by

which spatial variability in bedload flux develops, similar to such processes in large braided rivers. The bedforms in the miniature braided system are often obliterated in the dry ephemeral channel between flow events, giving rise to clockwise hysteretic bedload response due to bar reformation and the reorganization of the bed surface.

5. The sediment texture of the channel presents differences as large as one order of magnitude between the anabranch subsurface and the bar surface. Bedload texture is thought to vary depending on the topography of the bed. The GSD of the entire sampled bedload is similar to that of the bulk channel subsurface, implying that, on average, bedload transport is generally of equal mobility also when the segregated and unstable bar surfaces are mobile.

6. Measuring bedload in steep channels is a challenge also when the texture is sandy, as it develops a braided pattern. For future studies, it is recommended to accompany the monitoring of bedload with spatially distributed channel change data and simultaneous and accurate water discharge measurements to calculate hydraulic parameters such as stream power, average and local shear stress, thereby furthering our understanding of relevant morphodynamic processes and comparing them to those in other studied braided rivers.

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904

905 **Table captions**

906

907

908 Table 1: Summary of the monitored bedload-generating flow events (June 2009 -

909 January 2010) in the Barranca de los Pinos.

910

911 Table 2: Summary of spatio-temporal variations in bedload flux based on monitored

912 flow events in the Barranca de los Pinos (June 2009 - January 2010).

913

914 **Figure captions**

915

916

917 Fig.1. Location of the study area. The mesas and cuevas are capped by limestone and

918 dolostone (grey). The hillslopes, dissected by gullies, are underlain by horizontally-

919 bedded silica sand deposits, with thin intercalations of clay and gravel (black).

920

921 Fig. 2. Upstream view showing the two Reid-type bedload samplers in the Barranca de

922 los Pinos.

923

924 Fig. 3. Scatter graph of bedload flux vs shear stress for all bedload flux data in both

925 samplers (a); example of event (08/12/2010, event 23, right sampler) when bedload flux

926 is coherent with shear stress; $r^2 = 0.74$ (b).

927

928 Fig. 4. Example of clockwise hysteresis (direction of arrows) during event 19

929 (11/10/2010). LS= left sampler; RS= right sampler.

930

931 Fig. 5. Seasonality of the two kinds of temporal variation of bedload flux. Wave
932 occurrence varied seasonally less than did hysteresis, the latter was more frequent in
933 summer.

934 Fig. 6. Flow event 23 on 08/12/2010, when bedload flux in right sampler varied with
935 water depth – see Fig. 3b. (a). Flow event 6 on 19/01/2010 when bedload flux in left
936 sampler varied temporally while water depth remained essentially stable (b). Flow event
937 5 on 15/01/2010 when bedload flux in left sampler varied temporally in a wave-like
938 manner during quasi-constant increase in water depth (c).

939
940 Fig. 7. Example of bedload-generating flow event 10 (16/03/2010) when considerable
941 spatial variation in bedload flux occurs. Water is initially very shallow, supplying
942 bedload only to the right sampler. Overcoming a threshold in water depth, bedload is
943 thereafter also transported on the left side of the channel.

944
945 Fig. 8. D_{50} and D_{90} vs $\tau - \tau_c$ (shear stress minus the critical shear stress in both samplers
946 and averaged for 5 N m^{-2} bins. The critical shear stress was calculated using the Meyer
947 –Peter and Müller (1948) non-dimensional critical shear stress (0.047).

948
949 Fig. 9. Detail of the miniature braided pattern of the Barranca channel soon after a
950 bedload-generating flow event occurred (a); detail of the coarser bars and finer-grained
951 anabranches (b); high resolution DEM ($0.4 \times 0.4 \text{ mm}$) of a channel reach after
952 detrending the longitudinal slope (c), showing the braided pattern, the individual
953 bedforms (complex bars with chutes developing on top of them anabranches
954 surrounding them), and the location of cross sectional profiles along the braidplain
955 dominated by bars and anabranches (marked with arrows) (d).

956

957 Fig. 10. Texture of various riverbed units, with differences of one order of magnitude in
958 the texture of anabranch and bar surfaces (a). Grain size distributions of bedload
959 samples (thin light grey); the bold dashed black line represents the average anabranch
960 subsurface and the bold black line the bar surface. Bedload texture is on average well
961 represented by the average anabranch subsurface. Individual bedload samples are
962 considerably finer-grained than the subsurface and few others considerably coarser,
963 approaching that of the surface of bars. Time-weighted mean GSD of bedload (RS=
964 right sampler; LS= left sampler), is almost identical to the average channel GSD (b).

965

966 Fig. 11. Ratio of calculated/measured 10-min averaged bedload flux (i_B) in both
967 samplers. Bars indicate ranges; boxes indicate the 25 and 75 centiles; the median is
968 represented by a dash in the boxes.

969

970 Fig. 12. Temporal variation of bedload flux (right-sampler) and water depth during
971 events when monitoring duration exceeded 100 min.

972

973 Fig. 13. Evaluating the presence of a dominant frequency of bedload flux waves based
974 on Fourier analysis of bedload data for events when monitoring duration exceeded 100
975 min; water depth signal (a) and bedload flux signal (b).

976

977 **Tables:**

Table 1

event	sampling date	max. water depth	max. water depth before the sampler filled	max. shear stress	max. cumulative mass (LS)	max. cumulative mass (RS)	max. bedload flux (LS)	max. bedload flux (RS)	time gap: beginning of flow & bedload (LS)	time gap: beginning of flow & bedload (RS)
		mm	mm	N m ⁻²	kg	kg	kg s ⁻¹ m ⁻¹	kg s ⁻¹ m ⁻¹	min	min
1	01/10/2009	7	7	4.9	9.3	3.7	0.2	0.1	20.5	23
2	22/10/2009	13	13	9.1	22.0	5.1	0.8	0.1	2.5	3.5
3	02/12/2009	68	68	47.7	120.8	156.6	1.1	1.5	1.5	1.5
4	23/12/2009	134	134	93.6	88.8	118.4	13.8	1.5	11.7	7.5
5	15/01/2010	110	110	76.8	178.1	151.4	5.6	3.5	5	1.5
6	19/01/2010	50	48	35.6	173.3	-	2.1	-	42.5	-
7	05/02/2010	23	23	16.0	0	134.1	0	1.2	-	176
8	19/02/2010	69	69	48.9	180.3	169.2	3	1.5	10	4
10	16/03/2010	77	77	54.1	85.8	145.8	1.7	1.5	366	31.5
11	16/04/2010	40	40	28.1	80.9	23.44	5.3	1.49	0.5	2
12	11/05/2010	60	60	42.6	183.6	194.8	10.4	5.9	1	1.5
13	14/05/2010	54	54	38.2	120.5	172.9	5.6	19.45	12	0.5
14	01/06/2010	47	47	33.2	37.5	168.9	2.5	3.6	5	0.5
15	15/06/2010	155	77	107.0	167.4	124.3	7.6	8.4	3	0.5
16	06/07/2010	57	57	40.4	56.6	137.1	5.8	4.2	1	1
17	06/09/2010	80	80	56.2	168	163	19.9	6.7	11.5	0.5
18	24/09/2010	155	155	107	199.8	199.9	25.4	19.3	0.5	0.5
19	11/10/2010	36	36	25.2	91.9	132.4	2.7	3.1	0.5	0.5
20	01/11/2010	50	50	35.4	166	172.5	4.4	2	19	4
21	12/11/2010	63	63	44.7	178.7	185.5	8.4	3.5	0.5	0
23	08/12/2010	61	61	43.3	192.2	186.1	3.6	2.8	77	2.5

- the sampler was filled since the previous event, therefore bedload was not monitoredLS left sampler; RS – right sampler

978

Fig. 2

Fig. 3

Fig. 4

Fig. 5

Fig. 6

Fig. 7

Fig. 8

Fig. 9

Fig. 12

Fig. 13

979

980 Bedload flux first ever continuously monitored in a natural steep sandy channel.

981 The bedload fluxes reported herein are among the highest observed in a natural setting.

982 The quasi-flat channel bed is actually braided, which produces variability in bedload
983 flux.

984 Bedload texture expresses bar-anabranch textural differentiation but also equal mobility.

985