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1 **A multi-spatiotemporal scale strategy to evaluate factors controlling**
2 **pebble mobility and its interactions with bedforms in a lowland gravel-bed river**

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15
16
17 **Highlights**

- 18 • Very low bedload distances for full mobility floods at the event scale
- 19 • Weak trapping effect of bedload by bars at both event and decadal scales
- 20 • Bedload distances are lower than counterpart bedforms' distance at the event scale
- 21 • Bedload velocity varies depending on the year, bedform type or injection method
- 22 • Tracers' mixing was incomplete 10 years after their injection at the bed surface

27 **Abstract**

28 This paper aims to analyse how and what controls the spatiotemporal variability in bedload mobility in
29 a low-energy gravel-bed meandering river, with a particular focus on the interactions between bedforms
30 (bar-riffle-pool units) and bedload transport distance.

31 We first address the influence of pre-existing morphology, especially bars, upon the event
32 transport distance, and examine to what extent the maintenance of bedforms requires the occurrence of
33 hydrological events able to move the bedload over distances at least equivalent to the length of two
34 successive identical geomorphic units. To do that, we tracked the displacement of PIT-tags particles 10
35 years after their injection and applied bedload formulas to reconstruct the event transport distance for a
36 32-year-long study period. Our results show that event bedload distances for the bed surface D_{50} could
37 not exceed an active channel width, this one being roughly equivalent to 0.3 times the median distance
38 between riffles and between pools. This has two main implications. First, the trapping effect by bars
39 would be rare, or perhaps even absent, at the scale of competent events for most particles located at a
40 distance greater than an active channel width from a bar. Second, because no event is able to transport
41 bedload over distances at least equal to the length of one riffle-pool unit, the maintenance of bedforms
42 and bars especially, would arise from successive minor changes.

43 Further, our data revealed important variability in the bedload velocity at a decadal scale
44 according to the year of injection, the type of geomorphic unit of injection or the method of injection
45 (i.e., cluster or cross-section). This highlights the long-term persistence effects in low-energy rivers
46 upon velocity estimation from where, when and how tracers are injected.

47 Lastly, we inferred the long-term bedload velocity from the PIT-tags data and applied bedload
48 formulas. Equal to $0.95\text{--}1.3\text{ km century}^{-1}$, it is nevertheless considered an overestimate because the
49 slowdown of tracers seeded at the bed surface, due to their progressive mixing in the active layer and
50 their trapping in bars, was still on-going.

51

52 **Keywords:** Tracer dispersion; bedload path length; low-energy meanders; geomorphic units; active
53 layer

54

55 **1. Introduction**

56 In coarse-bed rivers, the bedload is generally considered responsible for shaping much of the fluvial
57 landscape (Andrews and Nankervis, 1995; Knighton, 1998; Leopold, 1992; Whiting et al., 1999).
58 Identifying and characterising the modalities of bedload incipient motion, displacements, and deposits
59 is therefore essential to improve our understanding of the conditions under which morphogenesis occurs.
60 Bedload movements are generally viewed as stochastic processes (Bradley and Tucker, 2012; Einstein,
61 1937; Paintal, 1971; Roseberry et al., 2012), these mainly controlled by three groups of parameters:
62 hydraulic and hydrological characteristics of the entrainment events, structural characteristics of the
63 sediments, and finally, their textural characteristics (Gilet et al., 2020; Hassan et al., 1991; Hassan and
64 Church, 1992; Vázquez-Tarrío and Batalla, 2019). Yet the distance travelled by the bedload during a
65 competent event does not always depend on stochastic processes alone (Pyrce and Ashmore, 2003a,
66 2003b). In addition to being influenced by the recent flow history and/or sediment supply (e.g.
67 Buffington and Montgomery, 1999; Church et al., 1998; Dietrich et al., 1989; Downs and Soar, 2021;
68 Hassan and Church, 2001; Hassan et al., 2020; Haynes and Pender, 2007; Konrad et al., 2002; Lisle et
69 al., 1993, 2000; Mao, 2012; Masteller et al., 2019; May and Prior, 2014; Monteith and Pender, 2005;
70 Nelson et al., 2009; Ockelford and Haynes, 2013; Paphitis and Collins, 2005; Reid et al., 1985), both
71 bed mobility and bedload distance could also be governed and constrained by a bed's morphology; i.e.,
72 its organization in well-structured geomorphic units, or bedforms. Accordingly, many researchers have
73 hypothesized that event transport distances scale with certain metrics that characterize bed morphology.
74 These are considered equivalent to the distance between an erosion zone and the closest downstream
75 deposit zone (e.g., Ashmore and Church, 1998; Carson and Griffith, 1989; Eaton and Lapointe, 2001;
76 Gaeuman et al., 2003; Lane et al., 1995; Neill, 1971, 1987). This assumption has been extensively
77 explored using various approaches that aim to estimate the transported volumes by applying the inverse
78 method (Ashmore and Church, 1998; Church, 2006). Furthermore, the influence of bed morphology on
79 travel distance, real or presumed, is more or less explicit in numerous studies (e.g., Beechie 2001;
80 Hassan and Church, 1992; Hassan et al., 1991; Hassan and Bradley, 2017; Kasprak et al., 2015; Kondolf
81 and Matthews, 1986; Lamarre and Roy, 2008; McDowell and Hassan, 2020; McDowell et al., 2021;

82 McEwan, 2001; McQueen et al., 2021; Papangelakis and Hassan, 2016; Pyrcce and Ashmore, 2003a,
83 2003b, 2005; Schmidt and Gintz, 1995; Sear, 1996; Vázquez-Tarrío et al., 2019).

84 According to Pyrcce and Ashmore (2003a, 2003b), the distribution of transport distances can be
85 categorised into four types, as distinguished by the intensity and duration of bedload-mobilizing events.
86 The first two types (types A and B in Pyrcce and Ashmore, 2003a) correspond to non-morphogenic
87 events. In both, the vast majority of particles are not mobilized or only transported over very short
88 distances; bedload displacements are hardly, or not at all, controlled by the bedforms, and; shear stresses
89 are slightly higher than the critical stress value and transport is partial. The last two types (types C and
90 D) correspond to morphogenic events. In both, transport distances are less dependent on the duration
91 and magnitude of floods, and instead increasingly influenced by bed morphology; shear stresses exceed
92 greatly the critical stress value and the bed eventually becomes fully mobile. Thus, in the case of
93 competent events that are sufficiently long and intense, the modal displacement distance is only partially
94 determined by flood characteristics (duration or intensity). Most of the material is then deposited on
95 bars. For riffle-pool systems, transport distances are equivalent to the distance between a pool and the
96 downstream bar; i.e., the distance separating two successive identical geomorphological units (Pyrcce
97 and Ashmore, 2003a, 2003b, 2005). Finally, during the longest and/or most intense floods, transport
98 distances are mainly controlled by the characteristics of hydrological events (Pyrcce and Ashmore, 2003a,
99 2003b, 2005; Reid et al., 2007). These events are potentially able to either remodel or destroy the original
100 morphology of the bed, by removing or remodelling the bars—thereby in effect limiting their influence
101 on transport distances—or remobilizing particles after their initial deposition on bars. Although based
102 on many field studies and flume experiments, this conceptual model needs to be tested against a larger
103 range of rivers and conditions to further assess its validity (Pyrcce and Ashmore, 2003a). Alternatively,
104 Pyrcce and Ashmore (2003a, 2005) also argued that the development and maintenance of these bedforms,
105 especially bars, requires the occurrence of hydrological episodes able to transport the bedload over a
106 distance similar to that between two successive bars, equivalent to the length of a single riffle-pool unit.
107 The existence of such morphology is hypothesized to result from the “net outcome of individual particle
108 displacements from discrete erosion loci” (Pyrcce and Ashmore, 2003a). However, that would imply that
109 bedforms are shaped by hydrological events with sufficient power for full mobility to occur, which

110 raises the question of whether such morphogenesis processes actually occur in low-energy gravel-bed
111 rivers flowing in a temperate environment, notably those having coarse bedload particles. Indeed, in the
112 case of most gravel-bed rivers, and more so for low-energy rivers, the highest shear stresses attained are
113 often only slightly greater than the critical shear stress. According to earlier work by Parker et al. (1982),
114 the ratio between the Shields stress and the critical Shields stress in most gravel-bed rivers almost never
115 exceeds a value of 2–3. Nevertheless, full mobility may only occur when this ratio is > 2 (Wilcock and
116 McArdeell, 1993; Church and Hassan, 2002; Recking, 2010). Furthermore, in temperate environments,
117 the difference in energy between frequent and rare flood events is generally low (Baker, 1977; Kochel,
118 1988; Lewin, 1989; Magilligan, 1992). These diverse aspects imply that in such systems, at least for
119 some, full mobility would be rarely attained (Church, 2006; Haschenburger, 2017; Haschenburger and
120 Wilcock, 2003). Yet, paradoxically, many of these rivers have a clearly discernible riffle-pool structure.

121 Here, we mainly examine the two sides of the conceptual model of Pyrcce and Ashmore (2003a,
122 2003b), by conducting an in-depth case study of a low-energy gravel-bed river. We focus on the
123 interactions between transport distance and morphogenesis in a meandering reach of the Cher River in
124 France across different temporal scales. Because of this river's low energy, we formulated two principal
125 hypotheses, both related to the assumed low event transport distances. First, only those particles located
126 at very close distances to bars could become trapped in them. Second, bedforms would be maintained
127 without the need of transport events whose distance is at least equal to the length scale of the riffle-pool
128 units.

129 To that end, using PIT tags, annual and decadal bedload transport distances were measured
130 between 2010 and 2020 along a meander bend. The distance travelled by the bed surface D_{50} at either
131 time scale was compared to that estimated according to recent bedload formulas. The ones best able to
132 predict the tagged particle distance at both yearly and decadal scales were then used to determine the
133 event transport distance for the entire discharge record period (i.e., 1988–2020) and then infer the
134 centennial bedload distance. Next, the distances obtained at the event and decadal scales were compared
135 to the distance separating successive identical geomorphic units. Finally, the tracers' 10 years of data
136 was used to investigate (1) the variability in bed mobility and velocity according to year, type of
137 geomorphic unit, and method of tracers' injection, and; (2) the way in which tracers were mixed within

138 the active layer. A key assumption being that the river's low energy confers a long-term persistence of
139 the effect on bedload velocity estimation arising from where, when, and how tracers are seeded.

140

141 **2. Study site**

142 After leaving its source at 713 m a.s.l., the Cher River mainly flows through gorges across steep slopes
143 or in deep valley with a very narrow floodplain (upstream Cher in Figure 1). The 63-km-long upstream
144 reach of the river is located at the northwest end of the low-elevation mountainous Massif Central
145 (France), which mainly consists of crystalline and metamorphic rocks (Larue, 1981, 2011). Then, the
146 alluvial plain begins (Alluvial Cher in Figure 1), and the river becomes single thread with a meandering
147 pattern that developed over the last millennia (Vayssière et al., 2020). After first occupying the Tertiary
148 graben of Montluçon, running 45 km to the confluence with the Aumance River (Larue, 1981, 2011;
149 Simon-Coinçon et al., 2000), the river then crosses the sedimentary domain of the Parisian Basin, where
150 it joins the Loire River at a 38 m a.s.l.

151 At the reference gauging station for our study site (Figure 1), the mean annual discharge is 14.8
152 $\text{m}^3 \text{s}^{-1}$. The rainfall regime is typical of lowland rivers on the western facades of mid-latitude regions.
153 The minimum and maximum monthly discharges occur respectively in August and February. The
154 hydrological regime, particularly the low-flow regime, is partly affected by the Rochebut Dam, it located
155 45 km upstream of the study reach, whose construction was completed in 1909 (Figure 1). Initially built
156 exclusively for electricity production, the dam has since been relied on to provide low-water support for
157 several decades now.

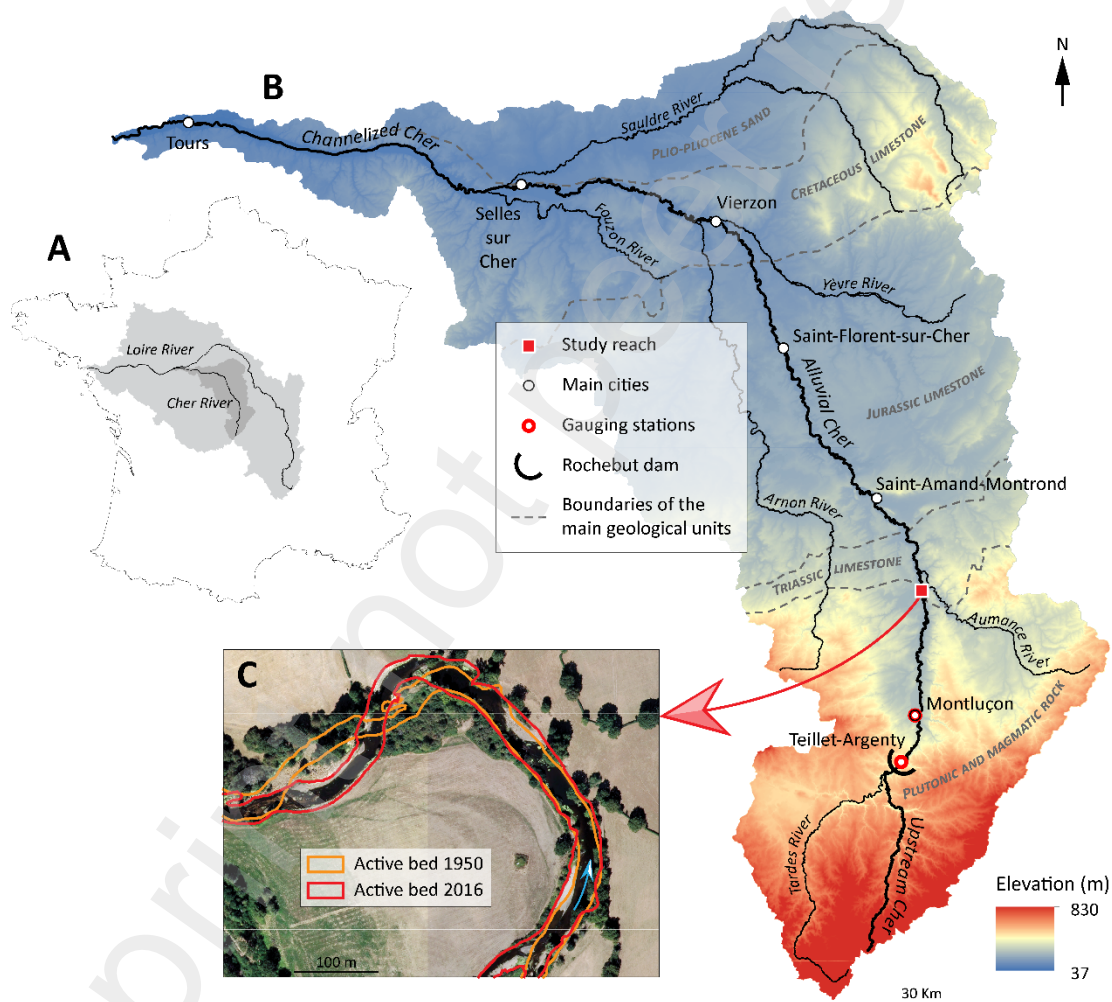
158 The study reach spans more than 700 m along a meander bend located in the Tertiary graben
159 of Montluçon (Figure 1). Its bedrock is composed of Eocene and Oligocene sand and clay on the left
160 side of the valley, and of Triassic sand more or less clayey on its right side (Turland et al., 1989a). The
161 alluvial infilling is schematically organized into two or three main stratigraphic levels. The base of the
162 infilling consists of sand, gravel, and pebble while its upper part consists of overflow deposits of clay,
163 silt, and sand (Dépret, 2014; Turland et al., 1989b); between these two levels, a predominantly sandy
164 layer is often interspersed. Because of this composite structure and the coarse nature of its material, the
165 banks are highly erodible (Dépret et al., 2015, 2017). But due to the presence of ancient fluvial

166 engineering works, especially those protecting the bank, the historical mobility of the reach is low
167 (Dépret et al., 2015, 2017).

168 The active channel width is 32 m, the low-flow bed slope is $0.001133 \text{ m m}^{-1}$, and the bankfull
169 discharge is $85 \text{ m}^3 \text{ s}^{-1}$ at the reference gauging station, this one located at Montluçon (Figure 1). With a
170 surface D_{50} and D_{84} of 28 and 48 mm, respectively, and a bankfull specific stream power of 30 W m^{-2}
171 (computed from the active channel width), it is a low-energy river with a coarse gravel bed (Dépret,
172 2014; Dépret et al., 2015). In the study bend, the armour ratio for the D_{50} and the D_{84} , as determined in
173 two riffles, is equal to 2.7–3.2 and 1.3–1.5, respectively (Dépret, 2014). These relatively low values are
174 consistent with the quite high frequency of mobilization of D_{50} (34 days y^{-1} for a critical discharge of 40
175 $\text{m}^3 \text{ s}^{-1}$) and the fact that the whole grain-size spectrum can be mobilized for discharge below the bankfull
176 (Dépret et al., 2017). These features strongly suggest a relatively weak sedimentary deficit, and thus, at
177 worst, a slight reduction in bed mobility because of the fluvial engineering works' presence or due to
178 the Rochebut Dam. An absence of the latter's influence is explained by the very lightly modified
179 hydrological regime for a discharge higher than the critical discharge, and by the fact that the total
180 interruption of the bedload continuity it has induced has not yet affected the study site. Comparing the
181 natural reconstructed daily discharge at the dam's outlet (powered discharge + variation of water level
182 in the reservoir + discharge above the crest of the dam; data provided by Electricité de France) with the
183 actual daily discharge outflowing from it (powered discharge + discharge above the crest of the dam;
184 data from <https://hydro.eaufrance.fr/>) shows that the annual number of days exceeding the critical
185 discharge is 1.2-times higher for the former than the latter. Concerning the peak of competent events,
186 the actual discharge values are just slightly lower than the natural reconstructed ones ($\text{ActualQ} = 1.04 \times$
187 $\text{NaturalQ} - 6.7$, $R^2 = 0.95$). Lastly, the distance between the Rochebut Dam and the study site is more
188 than 10-fold the distance travelled by bedload since the dam's completion. Hence, the bedload deficit
189 due to the dam could not have yet reached the study site.

190 The bedforms are organized into well-developed riffles and pools, some of them forced due to
191 the presence of the fluvial engineering works in the bed (Buffington et al., 2002; MacVicar et Roy,
192 2011; Thompson et Wohl, 2009). Therefore, the pools P1 and P2 can be separated in two sub-pools
193 (named A and B) according to the presence of a structure similar to a groyne (Figure 2). P3 is also

194 separated into two sub-pools (A and B) according to the presence of remnants of bank protections in the
 195 bed. P1B, P2B, and P3B are thus at least partly forced. Furthermore, the pool P3A as well as the riffle
 196 R3 are also probably forced given the presence of two crosswise, parallel alignments of piles on the
 197 right half of the bed, these exhumed due to bank erosion over recent decades (Figure 2). Furthermore,
 198 some bars are evident. The two main ones are situated at the upstream and downstream boundaries of
 199 the bend equipped with tracers (Figure 2). The downstream one (B4) is located along the study reach's
 200 main section, where lateral mobility has occurred in the last decades. For this reason, it has grown
 201 laterally during the study period, whereas the other bars have remained more stable.



202
 203 Figure 1 – Study site localization. A: The Cher River watershed within French territory. B: The study
 204 reach in the Cher River watershed. C: Zoom on the study reach.

205
 206

207 3. Material and methods

208 3.1 PIT-tags and bedload velocity from 2010 to 2020

209 Bedload velocity was estimated at an annual and decadal scale on a meander bend by tracking particles
210 equipped with passive transponders (Texas Instruments RFID chips, length: 12–23 mm, emission
211 frequency: 134.2 kHz). A total of 1503 tagged particles were injected in 2010, 2011, and 2012 into the
212 different geomorphological units of the bend (i.e., riffles, pools, bars) as well as into the two active
213 banks and at their toe (Figure 2). Previously, the geomorphic units were identified visually and
214 delineated with an RTK-DGPS: we distinguished riffles, pools, bars and also the main glides or runs.
215 Portions of some pools were also apparently forced (see section 2). Two injection strategies were used
216 in tandem, whereby tracers were introduced individually or in clusters. Individual tracers were injected
217 in 2010 and 2012 at 50-cm intervals along cross-sections located at the head of riffles. The total number
218 of particles per cross-section ranged from 32 to 53. Clusters were injected in 2010 and 2011, with 40
219 particles per cluster. In 2010, they were deposited in the centre of each of riffle and likewise in two of
220 the pools (P1A, P2B), and also at the head of the main migrating bar (B4) and in the banks as well as at
221 their toe (Figure 2). An identical strategy was implemented for the 2011 clusters, this time with all the
222 pools and the upstream run equipped with them (Figure 2). Clusters at the bank toe were positioned
223 midway between the upstream and downstream limits of the active banks. Clusters in banks were
224 positioned at three longitudinal locations: at the vertical of the clusters located at bank toe, and midway
225 between that point and the upstream or downstream limit of the active banks. Two clusters per vertical
226 were placed in the upstream bank (at mid-height and between that and the top of the bank), one in the
227 downstream bank (at mid-height). Except for the deepest pools (i.e., P1A, P1B, P2B, cf. Figure 2), all
228 the tagged particles were inserted into the pre-existing sedimentary structure, in order to avoid artificial
229 overexposure to the flows. For riffles, the particle size of tagged particles corresponds to the channel's
230 surface particle size (from 22.6 mm), as determined in 2010 by sampling 400 particles according to the
231 Wolman's methodology (1954). For the other geomorphic units, the tracers' particle size distribution
232 was the same as that for the riffle located immediately upstream. Each half-phi size-class was tagged
233 between 22.6 and 90.5 mm. For banks, only the 22.3–32 mm and the 32–45 mm classes were equipped.
234 Three detection surveys were carried out in summer at low flow, one each in 2011, 2012, and 2020

235 (detection system built by CIPAM). The detection front was located 180 m from the downstream-
236 injection cluster in 2011 and 2012 (for a maximum and a 99% percentile of particle distances of 232 m
237 and 117 m, respectively, for the 2010–2012 period). In 2020, the detection front was located 1200 m
238 from the downstream-injection cluster (maximum and 99% percentile of particle distances of 1200 m
239 and 720 m, respectively from 2010/2011/2012 to 2020). The position of each particle was surveyed
240 manually with an RTK-DGPS in the wadeable portions of the bed. For the rest of the bed, particles'
241 detection required the use of a boat. In 2011 and 2012, the boat was attached to ropes stretched across
242 the bed; in 2020, their detection was performed with an electric motor boat, the RFID reader and the
243 RTK-DGPS being connected to a laptop, with the position and identification of every tracer
244 automatically recorded (ArpenGIS software).

245 For particles individually injected along the cross-sections, the critical distance of mobilization
246 was set to 40 cm, which corresponds roughly to the maximum detection range of the antenna (Arnaud
247 et al., 2015; Chapuis et al., 2014). For clusters, the determination of this distance took into account the
248 uncertainty related to the initial position of each particle, as well as any uncertainty concerning the
249 location of the particles at the time of their detection. The exact initial position of each particle could
250 not be precisely known because we only had the coordinates for the centre of each cluster. Therefore,
251 the value of the first type of uncertainty is equal to the radius of clusters (40 cm in 2010, and 30 cm in
252 2011). The second type of uncertainty is linked to the maximum detection range (20 cm) of the small
253 antenna used to detect tracers located near the clusters. For most clusters, the total uncertainty, and hence
254 the critical distance of mobilization, was 60 cm (40 cm + 20 cm) for the 2010 particles and 50 cm (30
255 cm + 20 cm) for the 2011 particles. For some clusters located in pools, the critical distance was longer
256 because of greater uncertainty associated with the initial position and/or location of the particles (2 m
257 for M1A, P1B, and P2B; 0.8 m for P3A, P3B, and P4).

258

259 **3.2 Characterization of bedload mobility, velocity, and trajectory at a decadal scale**

260 For all particles, and likewise for each half-phi grain-size class, we first calculated the recovery rate, the
261 mobilization rate, as well as the mean and median distances travelled by the detected particles
262 (mobilized and non-mobilized). We also investigated how these parameters varied by date, by unit, and

263 by method of injection (cross-sections or clusters). To enable such comparisons, we distinguished those
264 particles injected in clusters into the same units in 2010 and 2011 (i.e., five riffles and two pools),
265 particles injected in clusters into riffles only in 2010 and 2011 (fives riffles), particles injected along
266 cross-sections in 2010 and 2012 (four riffles), particles injected in clusters into pools in 2010 and 2011
267 (two pools in 2010, seven in 2011).

268 Next, we examined the distances and trajectories of particles, to determine how they interacted
269 with the river's size and its geomorphic units, especially bars. To do that, we first calculated the
270 percentage of particles (mobilized and non-mobilized) that were deposited in their unit of injection or
271 in the downstream one. Then we compared the 10-year distance travelled by particles (= 10 times the
272 annual velocity of 2010 or 2011 or 2012 particles between their year of injection and 2020) to the median
273 distance between successive riffles and between successive pools (110 m); the percentage of travelled
274 distances lower than the median distance between units was determined. We also characterized the
275 influence of the only laterally developing and migrating bar (B4) upon the trapping efficiency of
276 particles at a decadal scale. For each geomorphic unit of tag injection upstream of the bar (all dates,
277 clusters, and cross-sections taken together), we looked at the percentage of all particles and those of
278 22.3–32 mm in size that got trapped in the bar and examined how it varied with distance to the bar.
279 Finally, we focused on particles injected in banks and at the bank toe, to better understand how particles
280 eroded from the banks are redistributed in the active channel.

281

282 **3.3 Distribution of particle velocities at a decadal scale**

283 The shape of the velocity distribution and of its tail was examined, to gain insight into the dispersion of
284 tracers occurs and because it has major implications for modelling advection-diffusion processes
285 (Haschenburger, 2013; Hassan et al., 2013; Hassan and Bradley, 2017). First, we determined how well
286 (or not) various probability distributions widely tested in the literature fitted with the annual velocity
287 distribution (for all generations of tagged particles taken together and for each generation considered
288 individually, for all particle sizes and only those 22.3–32 mm in size). Following the probability
289 distributions tested by Haschenburger (2013) and Papangelakis and Hassan (2016), and according to the
290 shapes of the annual velocity distributions, we tested four types of probability distributions using the

291 'fitdistrplus' package in R (Delignette-Muller and Dutang, 2015): the exponential, gamma, Pareto, and
 292 Weibull. Akaike's Information Criterion (AIC) was used as goodness-of-fit criterion between the
 293 probability distributions (Delignette-Muller and Dutang, 2015). Second, we focused on the behaviour
 294 of the distributions' tail. To do this, we implemented two approaches developed by Hassan et al. (2013).
 295 In one, a power law is fitted to the tail of the exceedance probability distribution of tracer velocities,
 296 using ordinary least-squares regression, and the fitted power value examined. Heavy and thin tails
 297 respectively correspond to a log-log slope of the exceedance tail of $\alpha < 2$ and $\alpha > 2$. The other
 298 approach relies on Hill's estimators (Hill, 1975), a ranking method based on the conditional maximum
 299 likelihood estimator for the power and slope in the Pareto distribution $P(X > x) = Cx^{-\alpha}$ fitted to the
 300 tail of sample data:

$$301 \quad \alpha_H = r \left[\sum_{i=1}^r (\ln X_i - \ln X_{(r+1)}) \right]^{-1}; C_H = \frac{r}{n} (X_{(r+1)})^{\alpha_H}$$

302 where α_H and C_H are estimators for power and slope, respectively; X_i are the order statistics such that X_1
 303 $\geq X_2 \geq \dots X_n$; the r is the rank of the smallest data point in the tail of the distribution. When $\alpha_H < 2$,
 304 the distribution is considered to be heavy-tailed.

305

306 **3.4 Estimation of bedload distance at the event-based and century scales**

307 To estimate the bedload velocity at the event-based scale, we tested the most recent formulas published
 308 in the literature (eq. 4 in Gilet et al., 2020; eq. 13, 14 for all rivers, eq. 14 for riffle-pool rivers, and eq.
 309 15 for riffle-pool rivers in Vázquez-Tarrío and Batalla, 2019; eq. in figure 2 for riffle-pool rivers with
 310 width/depth bankfull ratio > 10 in Vázquez-Tarrío et al., 2019; eq. for riffle-pool rivers in figure 4 in
 311 Vázquez-Tarrío et al., 2019; eq. for all rivers except for the unconstrained step-pool rivers in Figure 9
 312 in Vázquez-Tarrío et al., 2019). Derived from field data, these formulas relate the event bedload velocity
 313 or distance to different controlling parameters, such as flood magnitude and duration, grain-size of the
 314 bed, and the width and slope of the river, among others. Here we selected the ones providing the closest
 315 estimate of empirical distances travelled by the PIT-tags in the periods 2010–2011, 2011–2012, 2010–
 316 2020, 2011–2020, and 2012–2020. Here we postulated that if the formulas are robust for reconstructing
 317 the cumulative event bedload distance at yearly and decadal scales, they should be reliable for estimating

318 the event bedload distances. On this basis, the chosen formulas were applied to each competent event
319 that occurred for the entire daily discharge record period (i.e., 1988–2020 at the Montluçon reference
320 gauging station). The critical discharge for incipient motion was first estimated from presence–absence
321 PIT-tags surveyed at different discharge values between 2010 and 2012 along the cross-sections located
322 at riffle heads (Dépret et al., 2015, 2017) (R1, R3, R4, R5, cf. Figure 2). It was also estimated from six
323 20-mm sensitive ‘Benson-type’ seismic impact plate sensors (Downs et al., 2016), these equally spaced
324 along a cross-section located at the riffle head on the bend downstream of the PIT-tag bend, which
325 recorded data during the 2018–2019 high-flow period. Both approaches provided an estimation of the
326 critical discharge equal to $40 \text{ m}^3 \text{ s}^{-1}$ at the Montluçon reference gauging station. In addition to the critical
327 discharge, to apply the formulas three other parameters must be known: the surface grain-size, the bed
328 slope, and the active channel width. Surface grain-size was determined from low-flow photographs in
329 2018 of coarse grain-size patches of each bar. The photographs were taken vertically with the scene
330 hidden from direct solar radiation, at a high enough resolution to distinguish particles as small as coarse
331 sand. Using GIS (ArcGIS 10.6), we measured the apparent b-axis of 100 particles located at intersections
332 of a regular grid on the photo, whose spacing was higher than the maximum longest axis (a-axis). The
333 D_{50} and D_{84} used in formulas corresponded to the interpolated D_{50} and D_{84} of bars (28–48 mm), being
334 very close to the values obtained from the Wolman samples on riffles in 2010 (25–44 mm). The slope
335 was surveyed at low flow with a RTK-DGPS in 2020. The active channel width was reconstructed from
336 a 2016 aerial orthophotograph (provided by the IGN: Institut Géographique National) analysed in GIS;
337 after the active channel’s digitization, its surface area was divided by the length of the study reach to
338 obtain the average width.

339

340 **3.5 Interactions between bedforms and bedload distance at the event scale**

341 For a channel-forming discharge, namely for conditions of full mobility, and in the presence of well-
342 developed riffle-pool morphology, the modal event bedload velocity should be controlled by the
343 distance separating a pool to the bar located immediately downstream, the latter trapping most of the
344 particles (Pyrce and Ashmore, 2003a, 2003b). Conversely, this implies that for such a discharge,
345 particles should be able to displace to a distance at least equal to that separating two successive identical

346 geomorphic units. To test where that applies to the Cher River, we compared the event distances to the
347 median distance between successive riffles and between successive pools (sub-pools considered here as
348 individual units). The conditions for full mobility are theoretically reached when the grain shear stress
349 exceeds twice its critical value (Wilcock and McArdeell, 1993; Church and Hassan, 2002; Recking,
350 2010). To determine if events with full mobility had occurred from 1988 to 2020, the shear stress value
351 for which full mobility begins was derived, using two slightly different methods. In the first one, we
352 followed Recking et al.'s (2016) formula for riffle-pool rivers, which estimates the Shields stress
353 corresponding to the transition from partial to full mobility:

$$\tau_m^* = (5S + 0.06) \left(\frac{D_{84}}{D_{50}} \right)^{4.4}$$

354
355 where S is the bed slope (in m m^{-1}); D_{50} and D_{84} respectively are the median and 84th percentile of the
356 surface grain-size distribution (in m).
357

358
359 In the second method, we calculated the Shields stress value for the critical discharge, this
360 corresponding to incipient partial transport, multiply it by two to determine the Shields stress for full
361 mobility. The critical Shields stress was computed for the D_{84} , since it allows approaching the grain
362 stress without any bias (Recking et al., 2016), as follows:

$$\tau_c^* = \frac{\tau}{(\rho_s - \rho)gD_{84}}$$

363
364 where ρ_s is the density of sediment, ρ is the water density, and τ is the critical bed shear hear stress (in
365 N m^{-2}), the last obtained this way:
366

$$\tau = \rho g S d$$

367
368
369 where g is gravity acceleration (9.81 m s^{-1}); d is the average flow depth (in m), computed as follows
370 from the approximation of Rickenmann and Recking (2011) proposed by Recking et al. (2016):
371

372

373
$$d = 0.015D_{84} \frac{q^{*2p}}{p^{2.5}}$$

374 where $p = 0.24$ if $q^* < 100$ or 0.31 otherwise (which is the case of our study reach), and $q^* =$
375 $q/w\sqrt{gSD_{84}^3}$, for which q is the water discharge (in $\text{m}^3 \text{s}^{-1}$).

376

377 For overbank flows (i.e., those $> 85 \text{ m}^3 \text{ s}^{-1}$), the discharge q used is the one for the prism over
378 the channel. The bed mobility is controlled by this discharge value rather than by the total discharge.
379 Following Dépret et al. (2015), we used the results of a hydraulic modelling done in 1998 (Direction
380 Départementale de l'Allier, 1998) for seven flood discharge values along a cross-section located at riffle
381 R1. From it, a power law was obtained to predict the discharge over the channel as a function of total
382 overbank discharge:

383

384
$$q = 9.872Q_{tot}^{0.487}$$

385

386 where Q_{tot} denotes the total overbank discharge (in $\text{m}^3 \text{ s}^{-1}$).

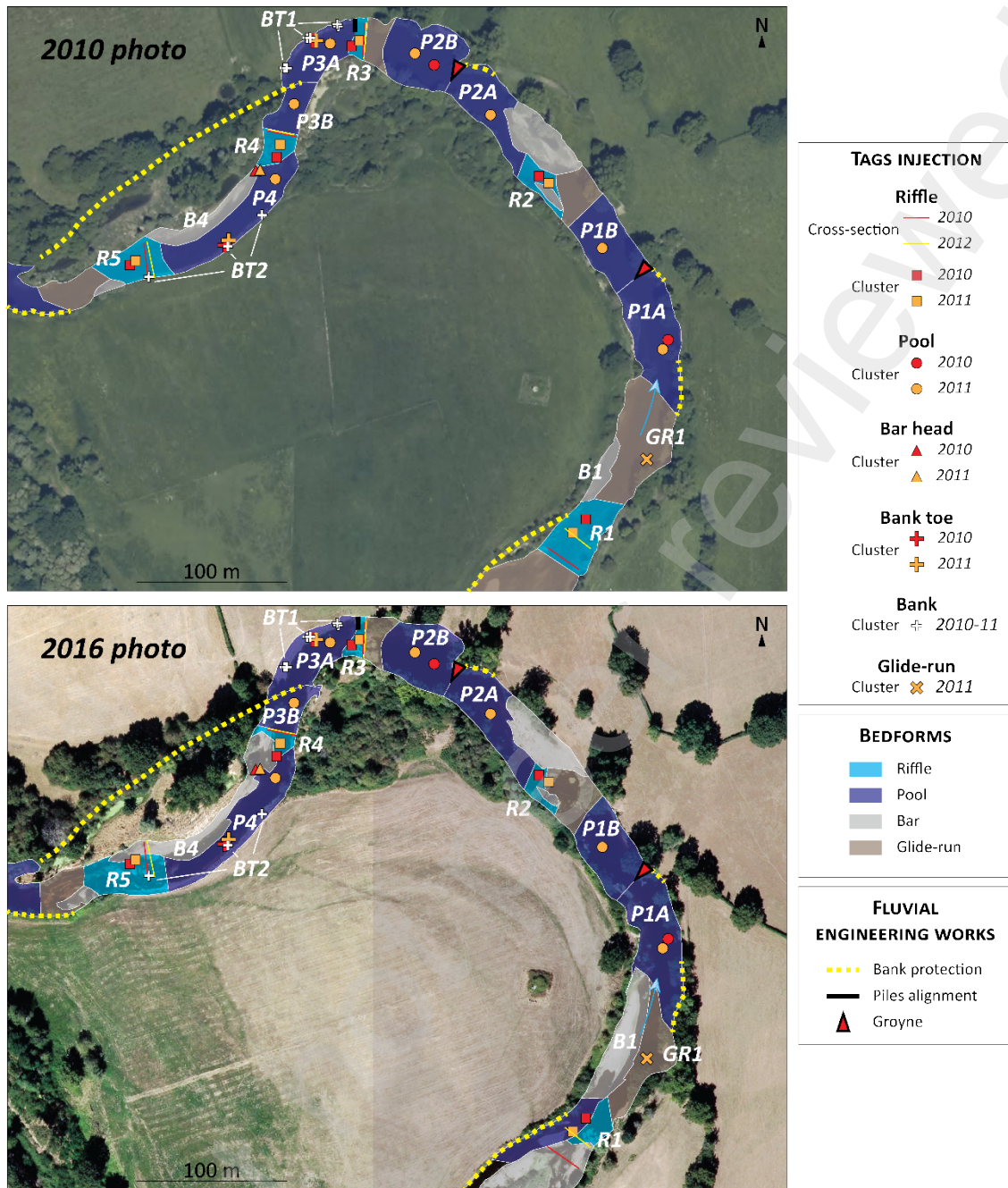
387

388 Then, according to both methods, we determined the discharge corresponding to the Shields
389 stress for full mobility. Each value was compared to the peak discharge of each competent event, to
390 identify those events in which the theoretically full mobility had occurred. Finally, for each event, we
391 compared the mean bedload distance to the median distance between riffles and between pools (110 m).

392

393

394



395

396 Figure 2 – Localization of the PIT-tags' injection and of the fluvial engineering works in the study reach;
 397 delineation of the main geomorphic units in 2010 and 2020.

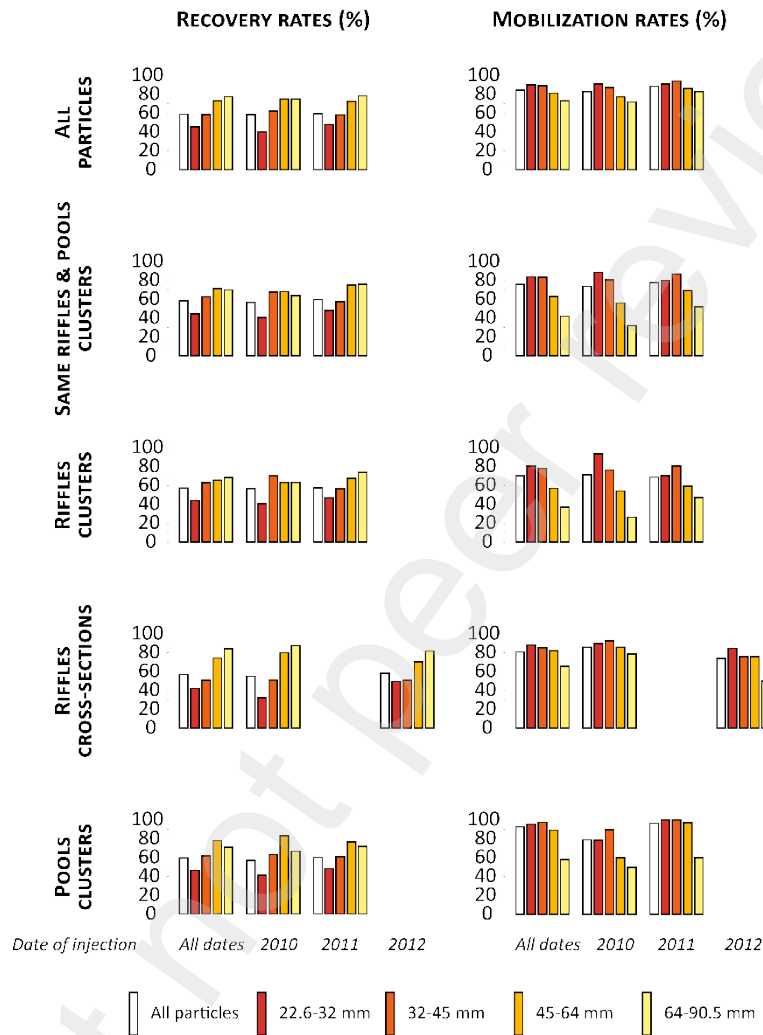
398

399 4. Results

400 4.1 2010-2020 recovery rate and mobilization rate

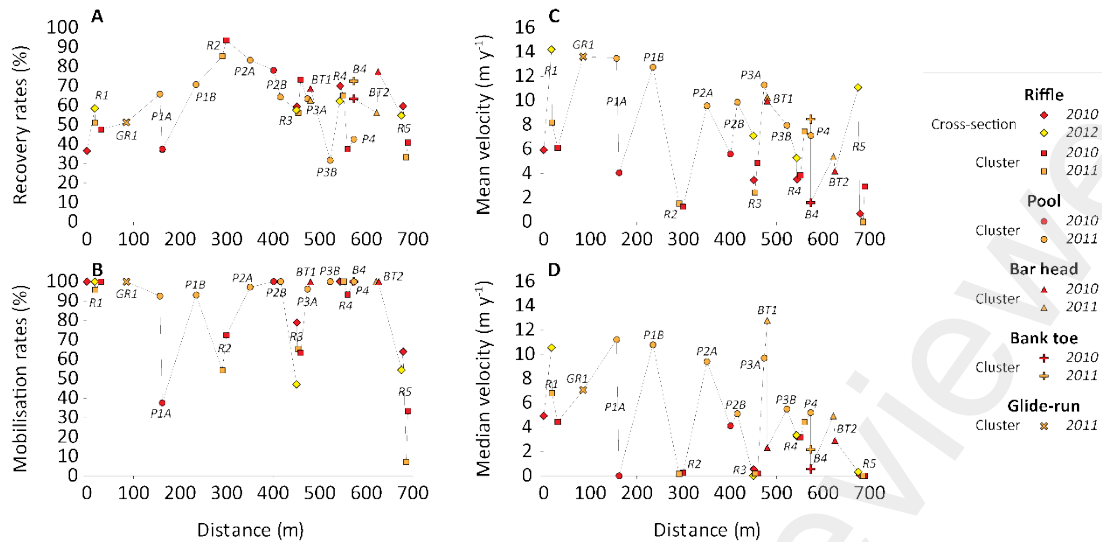
401 The recovery rate for all particles injected into the bed across all dates is 59% irrespective of injection
 402 year (Figure 3). Moreover, it increased with particle size: being 45% (40%– 49% according to the
 403 injection year), 59% (51%–62%), 73% (70%–75%, 78% (75%–82%) respectively for the particle size-

404 classes of 22.3–32 mm, 32–45 mm, 45–64 mm, and 64–90.5 mm (Figure 3). For particles injected into
 405 banks, the overall recovery rate was 53% (52% for injected particles in 2010, 55% for those in 2011).
 406 Considering individual injection points (cluster or cross-section), the recovery rates varied from 32% to
 407 100% (Figure 4A).



408

409 Figure 3 – Recovery rates and mobilization rates of tracers from 2010/2011/2012 to 2020.



410

411 Figure 4 – Longitudinal pattern of recovery rates (A), mobilisation rates (B), mean velocity (C), and
 412 median velocity (D) for all particle size-classes taken together.

413

414 The mobilization rate for all particles injected into the bed for all dates is 84% (83% for 2010 particles,
 415 89% for 2011 particles, and 73% for 2012 particles). It decreased slightly when the particle size
 416 increased from the 22.6–32 mm to the 45–64 mm size-class, but dropping sharply for the coarser class
 417 (64–90.5 mm). Rates were 90% (85%–91% according to the injected year), 89% (76%–94%), 81%
 418 (76%–86%), 73% (50%–83%) for particle size-classes of 22.3–32 mm, 32–45 mm, 45–64 mm, and 64–
 419 90.5 mm, respectively. For particles injected into the banks, the mobilization rate is 100% because of
 420 the bank retreat that occurred between 2010 and 2020. If we consider individually the injection points
 421 (cluster or cross-section), the mobilization rates varied from 7% to 100%, the second lowest rate being
 422 relatively higher at 33% (Figure 4B). Lastly, we observed a net difference in mobilization rates
 423 according to the year, the type of geomorphic unit, or even the injection method (cluster or cross-section)
 424 of tagged particles. The general pattern was of higher rates for 2011 particles, for particles injected in
 425 pools, and for particles injected along cross-sections (Figure 3, Figure 4B).

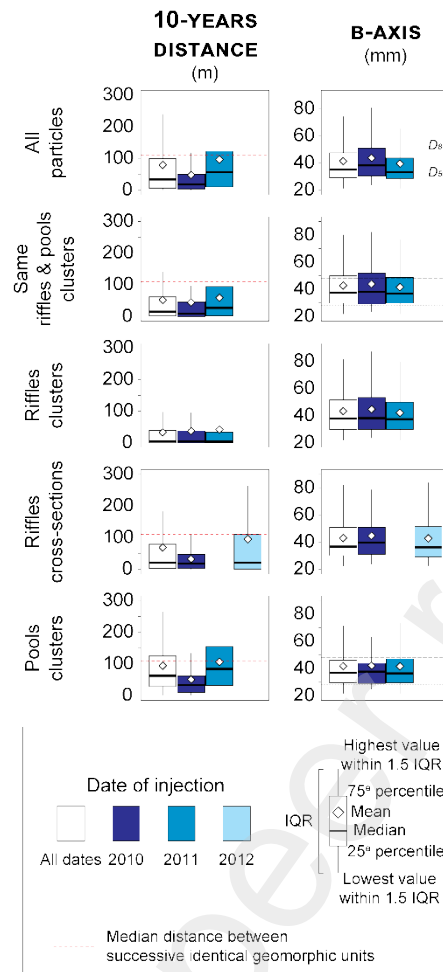
426

427 **4.2 2010-2020 bedload velocity**

428 The mean and median velocity for all particles injected in the bed for all dates are 6.8 m y⁻¹ and 2.9 m
 429 y⁻¹, respectively (Figure 5). Moreover, the velocity decreases with increasing grain-size: its mean

430 (median) is respectively 13 (7.1) m y⁻¹, 6.7 (4.6) m y⁻¹, 3.2 (1.6) m y⁻¹, and 1.1 (0.5) m y⁻¹ for particle
431 size-classes of 22.3–32 mm, 32–45 mm, 45–64 mm and 64–90.5 mm, respectively (Figure 6). In terms
432 of individual injection points in the bed (cluster or cross-section), the mean and median velocities varied
433 from 0.01 m y⁻¹ to 14.2 m y⁻¹ and from 0 m y⁻¹ to 12.8 m y⁻¹, respectively (Figure 4C and D). The
434 longitudinal pattern shows substantial variability but with slightly higher values along the upstream
435 stretch of 230–300 m (Figure 6).

436 Furthermore, not unlike the mobilization rates, we observed important variability in the bedload
437 velocity according to the year of injection, the type of geomorphic unit, and finally the method (cluster
438 or cross-section) used to inject the tagged particles (Figure 4C, Figure 4D, Figure 5, Figure 6). As these
439 observations are valid for all grain-size classes (taken individually or together), we go on to detail these
440 points solely for the 22.6–32 mm size-class. We did so for clarity of purpose and because the velocity
441 of particles belonging to this class and that of D₅₀ can be deemed reasonably similar: whichever way the
442 tagged particles are considered (by year or unit or method of injection), the mean and median grain-size
443 for the 22.6–32 mm size-class of the detected particles ranged between 27 and 30 mm, which is very
444 close to the bed surface D₅₀ on the studied bend (28 mm).



445

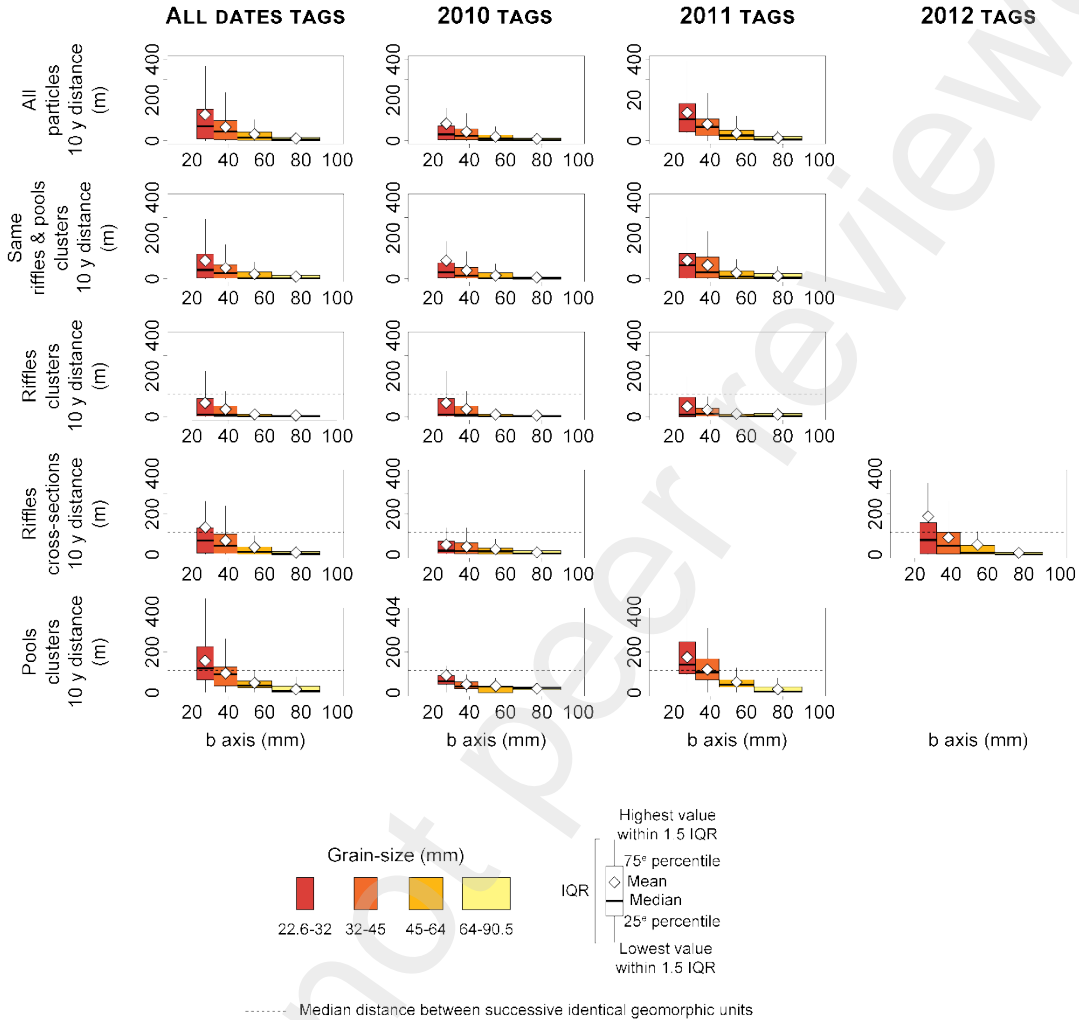
446 Figure 5 – Box-plots of 10-year distances and particle size for all particles grouped by date of injection,
 447 geomorphic unit and injection method.

448

449 First, the particles injected in 2010 were the slowest: their mean (median) velocity was 8.9 (3.1) m y⁻¹
 450 yet 9.5 (6.9) m y⁻¹ for the 2011 ones, when focusing on particles injected in clusters on the same riffles
 451 and pools in both years. For the particles injected individually along cross-sections in riffles in 2010 and
 452 2012, their velocity was 4.9 (2.1) and 18.8 (7.2) m y⁻¹, respectively (Figure 6). Second, the velocity of
 453 particles injected in pools was greater than that injected in riffles: for the 2010 particles, their velocity
 454 was 9 (1) m y⁻¹ for the former versus 8.6 (5.7) m y⁻¹ for the latter (Figure 6). The difference is more
 455 pronounced for the 2011 particles: 5.9 (1) m y⁻¹ for the riffle particles versus 16.9 (16.3) m y⁻¹ for the
 456 pool ones (when considering the same pools than in 2010, P1A and P2B; when considering all pools, as
 457 seen in Figure 6, the velocity is roughly the same: 17.5 (14.2) m y⁻¹). Lastly, particles injected along
 458 cross-sections moved faster than those injected in clusters (comparison only possible for riffles in 2010):

459 the mean (median) is 9 (1) m yr⁻¹ for the cluster particles and of 4.9 (2.1) m yr⁻¹ for the cross-section
 460 particles (Figure 6).

461



462

463 Figure 6 – Box-plots of 10-year distances for each half-phi class size.

464

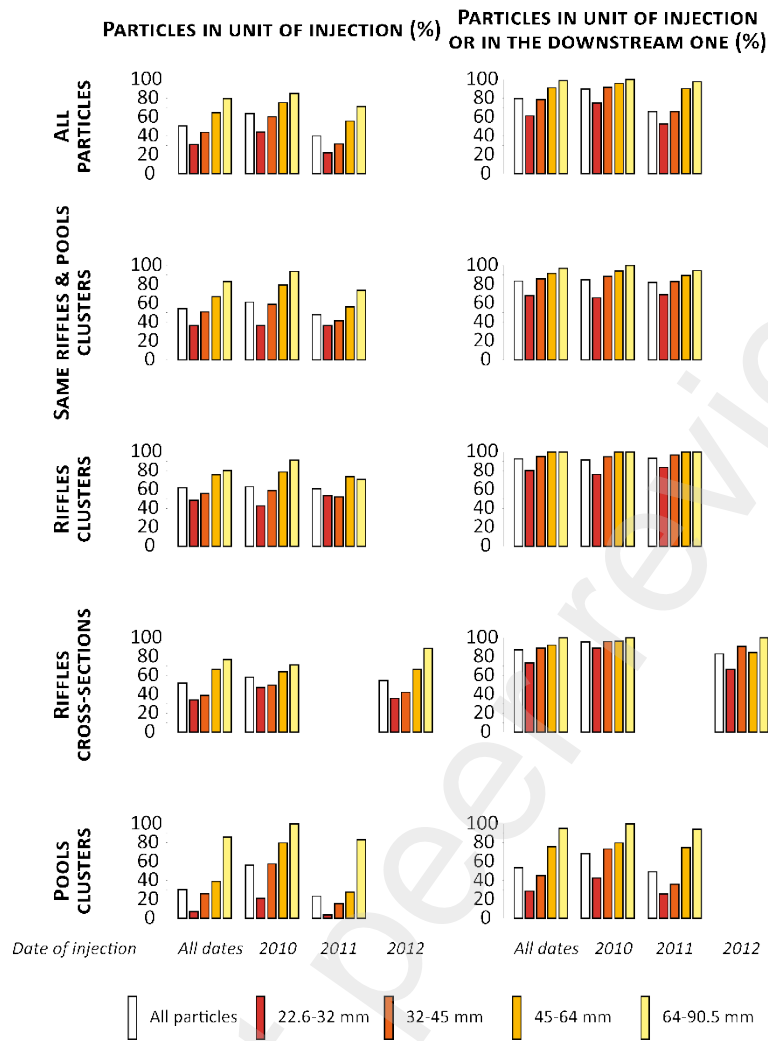
465 For the clusters in banks (BT1 and BT2, cf. Figure 2), the velocity was higher for the 2011 particles and
 466 those coming from the downstream bank. Specifically, for the only two size-classes equipped with tags,
 467 the mean (median) velocity was 18.8 (13.8) m yr⁻¹ for the 22.3–32 mm size-class but lower at 7.8 (3.6)
 468 m yr⁻¹ for the 32–45 mm size-class. The velocity for the 22.3–32 mm size-class was similar for the 2010
 469 and 2011 particles: 16.9 (14.2) versus 19.1 (13.7) m yr⁻¹, whereas for the 32–45 mm size-class, it is
 470 clearly higher in 2011: 9.7 (4.11) versus 5.2 (2.5) m yr⁻¹. Moreover, the velocity of particles was clearly
 471 lower for those injected into the upstream bank (BT1) than downstream bank (BT2): 13.2 (11.7) m yr⁻¹

472 versus 30.5 (16.2) m yr⁻¹ for all particles and all dates. In comparison to particles injected into banks,
473 the velocity of particles injected at the bank toe was lower for the 22.3–32 mm size-class (14.9 [7.2] m
474 yr⁻¹), but higher for the 32–45 mm size-class (8.6 [6.3] m yr⁻¹). Regarding the 2011 particles injected
475 into banks, they moved faster; however, the velocity of particles injected at the toe were higher for the
476 upstream bank. In detail, the velocity for the 22.32 mm size-class is 15.8 (4.4) m yr⁻¹ in 2010 and 13.9
477 (15) m yr⁻¹ in 2011. For the 32–45 mm size-class, it was 7.9 (5) m yr⁻¹ in 2010 and 9.4 (6.9) m yr⁻¹ in
478 2011. The velocity of 22.3–32 mm particles injected at the toe of the upstream bank was 23.8 (16.7) m
479 yr⁻¹. For the downstream bank, it was 8.2 (5.8) m yr⁻¹. For the 32–45 mm size-class, it was 12.7 (15.6)
480 m yr⁻¹ for upstream and 5.1 (4.9) m yr⁻¹ for downstream.

481

482 **4.3 2010-2020 bedload trajectory**

483 For the study period, half of particles injected into the bed (particles injected at bank toe and in banks
484 excluded) remained in their geomorphic unit of injection (Figure 7, Figure 8). On a per year basis, the
485 retained proportions were 64%, 40%, and 55% for the 2010, 2011, and 2012 particles, respectively
486 (Figure 7). The proportion of particles trapped in their geomorphic unit of injection increased with
487 particle size, being 31%, 44%, 65%, and 80% for the 22.3–32 mm, 32–45 mm, 45–64 mm, and 64–90.5
488 mm size-classes, respectively. The individual trajectory of particles can be seen for each size-class in
489 Figure 8. Furthermore, in 2020, the proportion of particles persisting in riffles surpassed that in pools
490 (34%–50% for riffles versus 7% for pools), despite much shorter distances from the injection point to
491 the exit units in riffles (mean, median of 21, 15 m for riffles versus 62, 69 m for pools) (Figure 7). For
492 the 22.6–32 mm size-class (which encompasses D₅₀), most particles initially injected into pools were
493 deposited outside of them (Figure 7). The difference in proportion between units decreased with particle
494 size. Lastly, 80% of particles injected into the bed (particles injected at bank toe and in banks excluded)
495 remained in their geomorphic unit of injection or in that located immediately downstream (Figure 7).
496 By year, the retained proportions were 90%, 72% and 83% for 2010, 2011 and 2012 particles
497 respectively (Figure 7). When examined by size-class, the proportions were 62%, 79%, 91%, and 99%
498 for the 22.3–32 mm, 32–45 mm, 45–64 mm, and 64–90.5 mm classes, respectively.

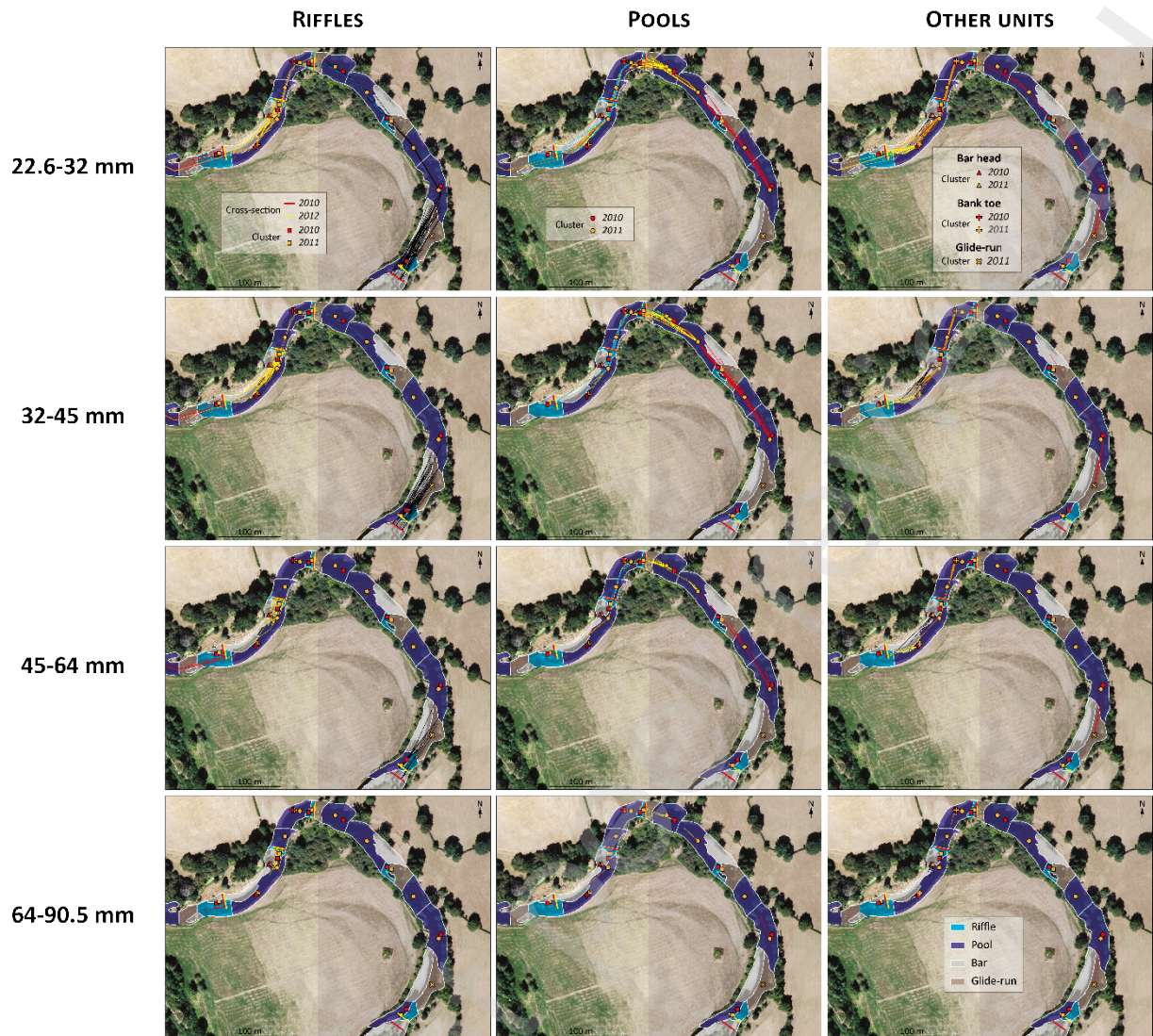


499

500 Figure 7 – Percentage of particles that remained in their unit of injection or deposited in the downstream

501 one in 2020.

502



503

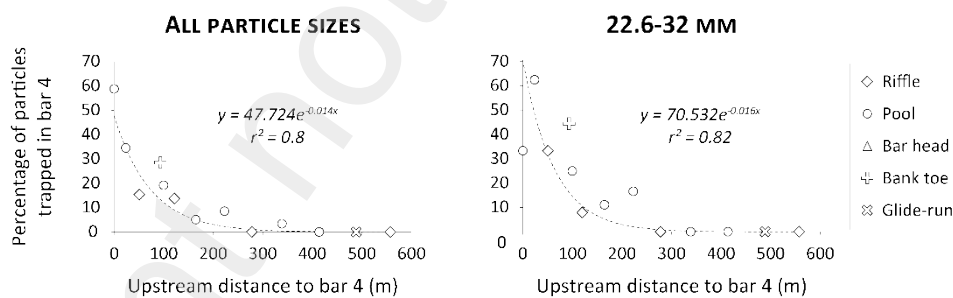
504 Figure 8 – Trajectory of individual tagged particles from 2010/2011/2012 to 2020.

505

506 The question of bedload trajectories was also addressed indirectly, by comparing the 10-year bedload
 507 velocity with the median distance between successive riffles and between successive pools (110 m). For
 508 all dates and particles injected into the bed, only 18% of particles travelled greater than the median
 509 distance between identical geomorphological units. By year, the proportions that did so were 7%, 23%,
 510 and 25% of the 2010, 2011, and 2012 particles, respectively; examined by size-class, it was 38%, 17%,
 511 6%, and 0% for the 22.3–32 mm, 32–45 mm, 45–64 mm, and 64–90.5 mm classes, respectively.
 512 Moreover, for all dates and particles injected into the bed, only 37% of particles travelled a distance
 513 longer than half the median distance between identical geomorphological units. Detailed by year, that
 514 proportion was 21%, 46%, and 35% for the 2010, 2011, and 2012 particles, respectively. Examined by

515 size-class, it was 59%, 45%, 18%, and 2% for the 22.3–32 mm, 32–45 mm, 45–64 mm, and 64–90.5
516 mm classes, respectively.

517 Finally, we also characterized the influence of the laterally migrating bar (B4) on the trapping
518 efficiency of particles at a decadal scale. For each geomorphic unit of tag injection upstream of the most
519 active bar (all dates, cluster and cross-section taken together), we looked at the percentage of particles
520 trapped in that bar. The Figure 9 shows that whether it be for all particles or only for the 22.3–32 mm
521 size-class, the percentage of particles trapped in the B4 bar strongly decreased along the 100–120 m
522 upstream of it; however, the influence of that bar on particle trapping then becomes null between 220
523 and 280 m upstream of it. Hence, the length scale of the bar B4's influence on bedload transport at a
524 decadal period seems to not exceed 220–280 m. Furthermore, most particles injected at the bar head
525 remained on the bar (92%). Finally, we looked at the percentage of 22.3–32 mm particles injected into
526 the upstream active bank and at its toe that were deposited in the B4 bar. For particles in the bank, this
527 bar trapped 36% of the tagged particles; for those at the bank toe, it trapped 44% of them. These values
528 exceeded somewhat those reported for the particles injected in the surrounding geomorphic units (8%
529 for S3, 25% for M3A, these located at the same cross-section as the cluster at the bank toe, and 33% for
530 M3B).



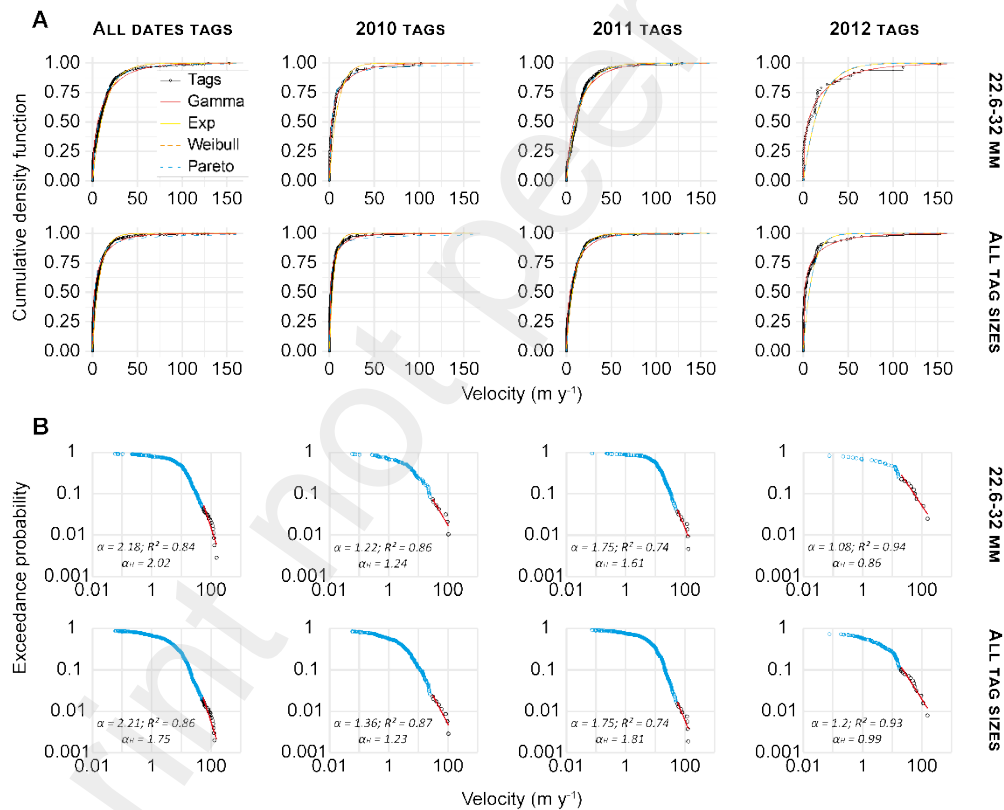
531
532 Figure 9 – Percentage of particles trapped in the bar B4 in 2020 in relation to the distance from the
533 injection point to the bar head.

534 535 4.4 Velocity distribution

536 Whether we examine particles all together or by date of injection, the distribution of their path length
537 was strongly right-skewed (Figure 10). The four probability distributions fitted—exponential, gamma,
538 Pareto, and Weibull—described well the annual velocity distribution (Figure 10A). Based on the AIC,

539 the best fit was obtained by the Pareto distribution (Pareto type 2), except for the 2012 tags, where
 540 gamma and exponential distributions appear better suited.

541 Finally, when the particles are differentiated by year of injection, the α and α_H values are
 542 consistently < 2 , indicating the strong likelihood of heavy-tail velocity distributions (for the 22.6–32
 543 size-class: α between 1.08 and 1.75 with an R^2 of 0.74–0.94, and range of α_H between 0.9 and 1.6; for
 544 all particle sizes: α between 1.12 and 1.75 with an R^2 of 0.74–0.93, and range of α_H between 0.99 and
 545 1.81) (Figure 10B). When all years of injection are taken together, the discrimination between thin and
 546 heavy tails was less clear in that most of α and α_H were slightly higher, even if relatively close, than 2
 547 ($\alpha = 2.12$, $R^2 = 0.86$, and $\alpha_H = 1.75$ for the 22.6–32 size-class; $\alpha = 2.18$, $R^2 = 0.84$, and $\alpha_H = 2.02$ for
 548 all particle sizes) (Figure 10B).

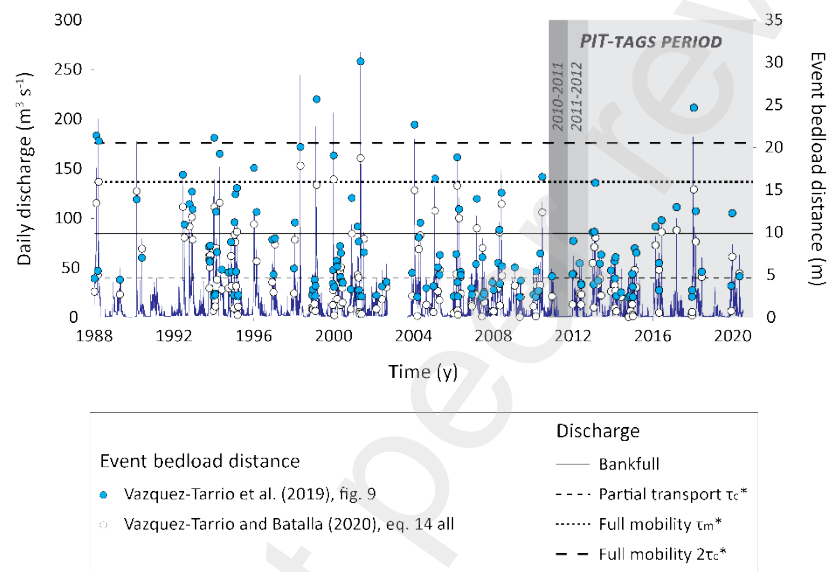


549 Figure 10 – Cumulative density function of annual velocities fitted with fitted exponential, gamma,
 550 Pareto, and Weibull distributions (A); Exceedance probability plots of annual velocities with α and α_H
 551 values of the tail’s distributions (tails are plotted in black) (B).
 552

553
 554

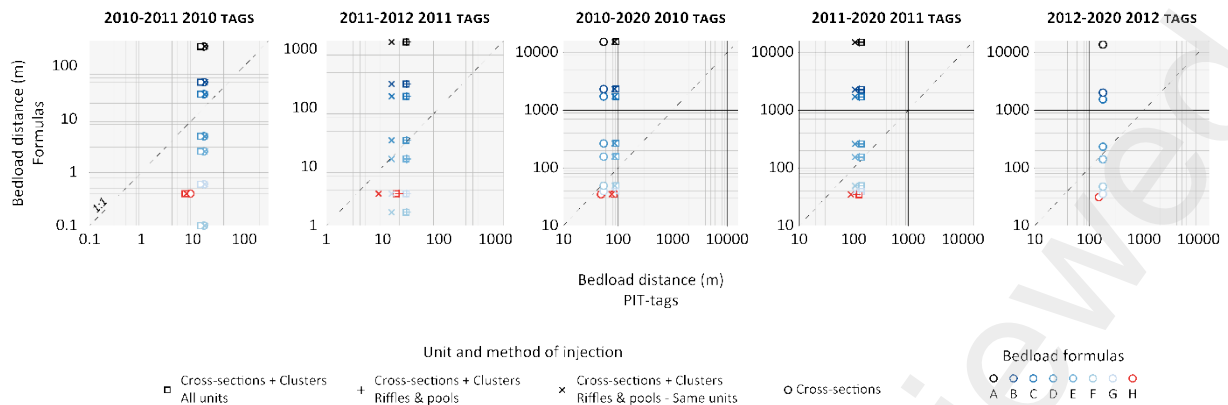
555 **4.5 Estimation of event and long-term bedload velocity**

556 For the whole discharge record (1988–2020) at the reference gauging station of Montluçon, 154
 557 competent events (50 events per decade) occurred, with a mean duration of 7 days per event (Figure
 558 11). Full mobility is theoretically reached at $137 \text{ m}^3 \text{ s}^{-1}$ and $176 \text{ m}^3 \text{ s}^{-1}$ (corresponding to return intervals
 559 of 3 years and 6 years, computed according Gumbel’s adjustment of annual maximum discharges) when
 560 estimated from τ_m^* and τ_c^* , respectively (Figure 11). This happened for 14 and 9 events from 1988 to
 561 2020, respectively (Figure 11).



562
 563 Figure 11 – Reconstruction of the mean event bedload distances for D_{50} from 1988 to 2020.

564
 565 Comparing the actual distance travelled by the bed D_{50} from tag tracking (22.6–32 mm size-
 566 class) to that predicted by recent formulas in the literature shows a relatively good agreement for those
 567 of figure 9 in Vázquez-Tarrio et al. (2019) (here, formula D in Figure 12) and for all types of
 568 rivers [eq. 14] in Vázquez-Tarrio and Batalla (2019) (here, formula D in Figure 12). This was verified
 569 for annual transport distance in 2010–2011, though to a lesser extent than in other periods, namely 2011–
 570 2012 and the longer periods of 2010–2020, 2011–2020, and 2012–2020 (Figure 12).



572

573 Figure 12 – Mean bedload distance (m) between 2010 and 2020 estimated using bedload formulas
 574 (with critical discharge values of $40 \text{ m}^3 \text{ s}^{-1}$ and $50 \text{ m}^3 \text{ s}^{-1}$) and PIT-tags tracking. The distances were
 575 obtained from mobilized and non-mobilized particles for all formulas, with the exception of formula H
 576 that computes distances from mobilized particles only. In bold, the distances obtained from formulas
 577 that are closest to the empirical tag distances. CSs: cross-sections. Cs: clusters.

578 A: Vázquez-Tarrío and Batalla (2019), eq. 13. B: Vázquez-Tarrío et al. (2019), eq. for riffle-pool rivers
 579 in figure 4 for riffle-pool rivers. C: Vázquez-Tarrío et al. (2019), eq. in figure 2 for riffle-pool rivers
 580 with width/depth bankfull ratio > 10 . D: Vázquez-Tarrío et al. (2019), eq. for all rivers except the step-
 581 pool ones of figure 9. E: Vázquez-Tarrío and Batalla (2019), eq. 14 for all rivers. F: Vázquez-Tarrío and
 582 Batalla (2019), eq. 14 for riffle-pool rivers. G: Vázquez-Tarrío and Batalla (2019), eq. 15 for riffle-pool
 583 rivers. H: Gilet et al. (2020), eq. 4.

584

585 These comparisons revealed that both formulas were well able to reconstruct the transport
 586 distances of D_{50} at the competent event scale on the studied bend. Between 1988 and 2020, the mean
 587 and median event distances for the 154 events that occurred were respectively 5.6 m and 7.7 m when
 588 using the formula D, and likewise 3.3 m and 5.1 m when using formula E. Of all events, 90% had a
 589 distance less than 12.2-15.7 m. The maximum distance was 18.8 m and 30.2 m under formula D and E,
 590 respectively, being equal or lower than the active channel width and 5.9–3.6 times lower than the median
 591 distance between riffles and between pools. The maximum distance attained was for a 31-day-long
 592 event, whose discharge was the highest during the study period (with a 34-year return interval period,
 593 computed according Gumbel's adjustment of annual maximum discharges) and for which the ratio

594 τ^*/τ_c^* was equal to 2.3. Not one of the 9–14 events for which full mobility occurred was thus able to
595 transport bedload on a distance equivalent to the distance between identical successive geomorphic
596 units. Even when applying the formulas to a fictional 10-day duration event with a constant 100-year
597 return interval discharge ($321 \text{ m}^3 \text{ s}^{-1}$) yielded very low distances: 22 m for formula D and 20 m for
598 formula E. Finally, extrapolating the cumulative distance obtained for the 1988–2020 period to a
599 duration of one century yielded an average long-term velocity of 2.5–3.8 km century⁻¹.

600

601 **5. Discussion**

602 **5.1 Variability in bed mobility and bedload velocity according to the year, type of** 603 **geomorphic unit, or method of injection**

604 Our results show there is important variability in both bed mobility and bedload velocity according to
605 the year, geomorphic unit type (and between units of the same type), and injection method (in cluster or
606 along cross-section) Understanding the reasons for such a discrepancy is a crucial issue in the context
607 of trying to obtain reliable bedload velocities.

608 Regarding the variability between particles injected in riffles versus pools, this could arise from
609 a difference in the available energy between these two types of units. According to the concept of
610 hydraulic reversibility, the average velocities and shear stresses during high flow should be higher in
611 pools (Booker et al., 2001; Caamano et al., 2009; Keller, 1971; Keller and Florsheim, 1993; Lisle, 1979;
612 MacWilliams Jr. et al., 2006; Milan et al., 2001; Petit, 1987; Thompson, 2011; Thompson et al., 1999).
613 Yet that process is not universal: it may only occur when the wetted cross-sectional area of riffles
614 exceeds that of pools (Carling, 1991; Clifford and Richards, 1992; Milan, 2013; Milan et al., 2002;
615 Thompson, 2011; Wilkinson et al., 2004), a situation rather infrequent (Carling and Wood, 1994;
616 Thompson, 2011; Thompson et al., 1999). For example, from a set of 702 pool-riffle couplets across
617 northern California, Byrne et al. (2021) found that reversal occurs at the bankfull for only 18% of all
618 couplets. The higher mobility of material in pools is perhaps more likely the result of the existence at
619 high-flow of convergent and jet flow at the entrance and centre of pools, respectively, and of divergent
620 and uniform flow at the pool tail and in riffles, respectively (e.g., MacWilliams Jr. et al., 2006;

621 Thompson and MacVicar, 2020). Furthermore, bed mobility and velocity could be promoted for tracers
622 injected into forced pools, namely for four of the seven pools equipped in our study (i.e., P1B, P2B,
623 P3A, P3B). In such units, the constriction imposed by the crosswise physical obstacles reinforced the
624 convergence of flows, and in such circumstances the flows near the bed surface are generally associated
625 with high turbulence (Hassan and Woodsmith, 2004; MacVicar and Roy, 2007a, 2007b, 2011;
626 Thompson, 2007; Thompson and Wohl, 2009). When tracers are injected in the deepest sections of the
627 forced pools, as done in our study, mobility is reportedly higher than in riffles (Thompson and Wohl,
628 2009; MacVicar and Roy, 2011). Yet our data do not show such an effect of forced pools: for each pair
629 of unforced–forced pools (M1A–M1B, M2A–M2B, M3A–M3B), the velocity was either higher or
630 similar for particles injected in the unforced pool. The difference between forced and free-formed pools
631 could nonetheless result from lower distances travelled through pools for particles injected in the former
632 (24 m, 38 m, and 21 m from the injection point to the exit of the forced pools M1B, M2B, and M3B,
633 respectively, versus 103 m, 87 m, and 69 m for the free-formed pools M1A, M2A, and M3A,
634 respectively). Furthermore, the lateral location of clusters can also explain, at least in part, the difference
635 in bed mobility and bedload velocity observed between riffles and pools. Whereas clusters in pools were
636 systematically located in their central deepest section, where the jet flow is supposed to occur, three
637 of the five riffle clusters (R2, R3, and R5) were located in the bend and not at the inflexion point (R1
638 and R4), where the clusters were positioned rather than in the inner half of the bed, where the flow
639 velocity could be relatively low. This explanation is supported by the clearly lower bed mobility and
640 bedload velocity of tracers injected in these three riffles compared with the two remaining riffles (range
641 of mobilization rate: 7%–72% for R2, R3, and R5 versus 93%–100% for R1 and R4; respective ranges
642 of mean, median velocity: 0.8–4.9 m y⁻¹, 0–0.3 m y⁻¹ versus 3.8–8.2, 0–0.3 m y⁻¹ and 3.2–6.8, 0–0.3 m
643 y⁻¹). When compared with pools, the mobilization rate as well as bedload velocity of these three riffles
644 (R2, R3, R5) are also clearly lower, but similar for the two remaining riffles. In addition,
645 sedimentological arguments can also be put forward. Some research demonstrated that the structure of
646 the surface layer of pools (their degree of packing especially) is clearly less stable than for riffles (e.g.,
647 Clifford, 1993; Hodge et al., 2013; Sear, 1996). Further, during low flow, the competence in riffles
648 surpasses that in pools (Milan et al., 2002): the finest fraction of the bedload, sand, and small gravel, is

649 at least partly exported from the former and accumulated in the latter (Jackson and Beschta, 1982; Lisle
650 and Hilton, 1992, 1999; Milan, 2013; Sear, 1996). The pools are thus frequently lined with a more or
651 less thick layer of fine material and the sand content in their subsurface layer could be more important
652 than in riffles. This may have been a factor contributing to greater bed mobility in pools (Sear, 1996;
653 Vahidi et al., 2020). Work by Curran and Wilcock (2005), Wilcock (1998) or Wilcock and Crowe
654 (2003), for example, show that significant increases in the mobility of gravel occur when the proportion
655 of sand in the bed increases. All the possible explanations presented above are theoretically valid at the
656 scale of the competent events, or so long as tracers remain in their unit of injection. Over time, given
657 the tracers' alternating passage through riffles and pools, the discrepancy in velocity should eventually
658 slow to zero. When considering only those tracers that left their unit of injection, we still found a marked
659 difference in bedload velocity between tracers seeded in riffles and in pools: for 2010 tracers, mean
660 (median) of 11 (5.6) m y^{-1} and of 9.6 (6.7) m y^{-1} for riffles and pools, respectively; for 2011 tracers,
661 mean-median of 8.1 (6.3) m y^{-1} and 13.7 (12.6) m y^{-1} for riffles and pools, respectively. This shows that
662 the effect on the bedload velocity from the type of unit that received the injection can persist 10 years
663 after the injection. This most likely arose from the low bedload velocity reported: half of the tracers
664 remained in their unit of injection (31% for the 22.6–32 mm size-class) and 80% of tracers were
665 deposited in their unit of injection or in the downstream one (62% for the 22.6–32 mm size-class) (Figure
666 7). Secondly, it is also probably reflects the effect from the difference in distance from injection points
667 to the exit of a geomorphic unit according to its type. This distance is clearly shorter in riffles than pools;
668 mean (median) of 21 (15) m versus 62 (69) m, and references back to the proportions leaving the two
669 types of units. However, lacking detailed data on the hydraulic functioning and surface sediment
670 structure of these units, which of the above potential explanations—for the difference in velocity
671 between particles injected in riffles versus pools—is prominent cannot be known.

672 The lower velocity of particles injected as clusters than along cross-sections likely originates
673 from the bed mobility's laterally discrete nature. For partial transport events, namely for most of the
674 competent events, only a portion of the bed cross-section, which expands with the flow strength, is
675 indeed mobile (Ferguson, 2003; Frings and Vollmer, 2017; Haschenburger, 2017; Haschebburger and
676 Wilcock, 2003; Papangelakis and Hassan, 2016). Injections performed in a cluster would thus provide

677 an under-estimation of the bedload velocity when the cluster is positioned on a rarely active portion.
678 Conversely, that velocity could be overestimated if the cluster is located on a frequently active part of
679 the bed. Broadly then, injecting along a cross-section would let us incorporate the lateral variability of
680 bed mobility and thus yield a more representative image of bedload velocity.

681 In order to integrate the effects of these different sources of variability for bed mobility—that is
682 laterally, between types of geomorphic units, as well as between units of the same type)—thereby obtain
683 the most reliable bedload velocity possible, we recommend using seed tracers in several equidistant
684 clusters along cross-sections, with a least two or three particles of each half-phi grain-size per cluster.
685 Such injections should be made in several riffles and in several pools, with cross-sections systematically
686 located at the same relative longitudinal position (i.e., heads of riffles and centres of pools, for example).
687 Whatever the chosen injection design, the variability in bedload velocity should lessen over time and be
688 paramount when full mobility is rare.

689 The inherent spatial variability of bed mobility and velocity could also explain, in part, the lower
690 mobility and velocity observed for 2010 tags vis-à-vis 2011 and 2012 to the extent that particles were
691 injected at a slightly different position in a given year versus others. Operator bias is very likely
692 negligible, in that particles were injected in the same way and by the same person. The last potential
693 explanation is related to the differing flow history between tracers: because 2010 and 2011 tracers share
694 the same hydrology following the injection of 2011 tracers, an explanation for the lowest decadal
695 velocity of the 2010 tracers could theoretically lie in what happened before the injection of the 2011
696 tracers. Indeed, the 2010 tracers underwent a very long low flow period prior to the 2011 tracers'
697 injection (Figure 11), that could generated a rearrangement of the surface structure that led to its
698 strengthening, thus limiting the particles' mobility. Numerous authors have highlighted the fundamental
699 role of stress history, that is to say sub-critical shear stresses, in strengthening the surface layer stability
700 (Haynes and Pender, 2007; Monteith and Pender, 2005; Ockelford and Haynes, 2013; Paphitis and
701 Collins, 2005). The effect of such limitations for the 2010 tracers' mobility should have emerged in the
702 2010–2020 period, and accordingly, in 2011–2012, the 2010 tracers ought to have moved slower than
703 the 2011 tracers. Yet the opposite occurred. When comparing tracers introduced in clusters to the same
704 units (pools or riffles) in both years, the mean velocity of the 22.6–32 mm size-class is 12.9 m y^{-1} and

705 5.4 m y⁻¹ for the 2010 and 2011 tracers, respectively. Further, whereas the total expenditure of flow
706 energy for a discharge higher than the critical discharge (*sensu* Haschenburger, 2011a) was 9.4 times
707 lower in 2010–2011 than 2011–2012, the mean virtual velocity (distance divided by the duration of
708 competent discharge) of the 22.6–32 mm size-class was clearly higher for the 2010 tags in 2010–2011
709 than the tags 2011 in 2011–2012 (69.4 mm h⁻¹ versus 7.3 mm h⁻¹). We conclude the difference in the
710 location of the tracers' injection between years is the only satisfactory explanation for the variability in
711 velocity between years.

712

713 **5.2 Interactions between event bedload distances and bedforms**

714 For the entire period of recorded discharge (1988–2020), despite the occurrence of 9–14 events for
715 which the full mobility was reached, the maximal estimated mean event distance for the D₅₀ was 30.2
716 m. This is roughly equivalent to the active channel width and 3.6 times less than the median distance
717 between riffles and between pools. The main implications for the Cher River, and perhaps more broadly
718 for temperate low-energy gravel-bed rivers, are twofold. First, event transport distances would be
719 always, or almost always, shorter than the distance between two successive identical geomorphic units.
720 It implies that trapping effect by bars would be rare, or even absent, at the scale of competent events for
721 most of particles located at a distance higher than an active channel width. Second, since no event is
722 able to transport bedload on distances at least equal to the length of a single riffle-pool unit, the
723 maintenance of bedforms, and especially that of bars, would thus become disconnected from the event
724 transport distances or would depend on successive minor changes. Therefore, the model proposed by
725 Pyrcce and Ashmore (2003) would not be universally valid. When computing the cumulative distance of
726 D₅₀ for events whose peak is higher or lower than the bankfull discharge, we see that events below the
727 bankfull are responsible for about half of the total distance for the 1988–2020 period. By setting the
728 threshold to a 2-year return interval flood, events with lower discharge than that are responsible for ca.
729 70% of the total distance for the 1988–2020 period. This agrees with and supplements the findings of
730 Dépret et al. (2015) along the same reach on which the study bend is located, showing that its
731 morphogenesis, from the perspective of planform mobility and effective discharge for the bedload, is
732 chiefly controlled by low-magnitude hydrological events (< 2 years RI).

733 Finally, the possibility remains that the study bend's current bed morphology could be
734 maintained by very rare floods and/or had been inherited from a past flood regime. Still, along those
735 scarce sections of the Cher River whose geometry has freely adjusted over the last decades (due to the
736 lack of lateral constraints), there exists riffle–pool–bar sequences with length scales of bedforms similar
737 to that in the study reach. This suggests the present morphology can be maintained under the current
738 flood regime, and without the occurrence of a very rare flood event.

739

740 **5.3 Long-term bedload estimation**

741 The range of long-term bedload velocity we proposed here is an overestimate for two key reasons. First,
742 the study bend's hydraulic parameters are not representative of the whole 12-km long reach, which ends
743 at the junction with the Aumance River (see Dépret et al., 2015, 2017 for details) and is where the bend
744 is located. Although their active channel width is similar, the low-flow bed slope is equal to 0.00063 m
745 m⁻¹ along the reach, versus the greater 0.0011 m m⁻¹ along the study bend, where the bankfull stream
746 power is thus almost two times higher (30 W m⁻² versus 17 W m⁻²). This implies the long-term velocity
747 should be derived instead from the hydraulic parameters of the reach. By doing so, the application of
748 formulas D and E using hydraulic parameters of the reach provides velocities of 0.95 and 1.3 km century⁻¹,
749 respectively (versus 3.8 and 2.5 km century⁻¹ for the study bend). Secondly, it is very likely that our
750 estimation takes into account, only partially, the progressive slowdown of the tracers resulting from their
751 mixing within the event active layer (*sensu* Church and Haschenburger, 2017), or their trapping in bars
752 or in the floodplain (Ferguson et al., 2002; Ferguson and Hoey, 2002; Haschenburger, 2011a, 2011b;
753 Houbrechts et al., 2015; Iwasaki et al., 2018; Pelosi et al., 2016). For example, in the Carnation Creek,
754 Haschenburger (2011a) reported that a 72% reduction in velocity after a few years, which reached then
755 an asymptotic value. In the Allt Dubhaig, Ferguson et al. (2002) found a 50% reduction in velocity when
756 comparing it 2 years and 8 years following the tracer injection. Here, long-term bedload velocity in the
757 Cher River was estimated through formulas that succeeded in reconstructing the tracers' velocity at a
758 10-year scale. Yet, because of the low transport distances reported, we find that only a small proportion
759 of all tracers could have been affected by the trapping effect of bars on this time scale. Such a weak
760 effect is also evident in the similar velocity of tracers deposited on and outside the bar B4: respectively

761 having a mean (median) of 8 (5.3) and 8.1 (3.8) m y^{-1} for all particles; likewise, 13.6 (8.7) and 14.7 (10)
762 m y^{-1} for the 22–32 mm size-class. Furthermore, especially because of relatively weak hydrologic
763 activity during the 10-year study period, coupled with river’s inherently low energy, it is likely that the
764 mixing process of tracers into the active layer was still incomplete at 10 years since their injection. In
765 this respect, Haschenburger (2011a) showed, for example, that large floods accelerate the rate at which
766 the vertical mixing of surface tracers occurs. Recently, Vázquez-Tarrío et al. (2021) highlighted that, at
767 the event scale, the ratio between the Shields stress for the peak flow and the critical Shields stress is
768 one of the main drivers of the active layer’s depth. By comparing the virtual velocity (i.e., total distance
769 divided by the total duration of competent events) of tracers 1 year and 10 years following their injection,
770 we could indirectly gain insight into the intensity of their mixing process, the idea being that the smaller
771 the difference, the weaker is the mixing. For the 2010 tracers, the mean virtual velocity of the 22.6–32
772 mm size-class was much faster in the 2010–2011 period than 2010–2020 period (74 mm h^{-1} versus 13
773 mm h^{-1} , a 82% decrease). Yet, when comparing the median velocity, the velocity is higher for the longest
774 period (5 mm h^{-1} versus 3 mm h^{-1} , a 148% increase). We find a similar tendency for the 2011 tracers,
775 but with a clearly smaller difference for the mean (24 mm h^{-1} for the 2011–2012 period versus 20 mm
776 h^{-1} for the 2011–2020 period, a 15% decrease; for the median, 8 mm h^{-1} versus 15 mm h^{-1} , a 198%
777 increase). Hence, it seems virtual velocity has not or only marginally decreased with elapsed time. This
778 would tend to confirm a weak mixing of tracers in the active layer. The right-skewed distribution of the
779 bedload velocity and its likely heavy-tail (Figure 10) further support this interpretation of a mixing
780 having yet not reached its equilibrium. Both features, which could serve as a signature of a super-
781 diffusive regime of the bedload (Lajeunesse et al., 2013), would stem from the fact that a crucial
782 proportion of them remains overly mobile so long as they have not been fully mixed into the active layer
783 (Pelosi et al., 2016; Pierce and Hassan, 2020; Wu et al., 2019; Zhang et al., 2012). On a longer time-
784 scale, after the mixing has been achieved, the dispersion regime of the tracers would tend to become
785 diffusive, rendering their distribution of velocities more normal, with their variance increasing linearly
786 with time and their velocity becoming asymptotic (Guillon, 2016; Lajeunesse et al., 2018; Wu et al.,
787 2019; Zhang et al., 2012). But there is no consensus yet concerning the anomalous or normal character
788 of the dispersion of tracers in the long-term. Accordingly, Pierce and Hassan (2020) suggested that on

789 long time-scale—a geomorphic range as they called it—the regime would remain anomalous, for which
790 superdiffusion or subdiffusion happens in a way that depends on the long-time decay of the exhumation
791 time distribution of buried particles.

792

793 **6. Conclusion**

794 This paper focused on characterizing the variability of bed mobility in a low-energy, gravel-bed
795 meandering river, with particular attention paid to the relationship between bed transport distances and
796 bed forms. Our combined analysis of 10-years of data for tracked PIT-tags and a 32-year reconstruction
797 of bedload distance events during which full mobility was reached on several occasions reveals a
798 substantial slowness of the bedload in the studied river. Our findings highlight two take-home messages.
799 First, the trapping effect by bars is rare, or even absent, at the scale of competent events for most of
800 particles located at a distance greater than an active channel width from the bar. Second, since no event
801 is able to transport bedload over distances at least equal to the length of one riffle-pool unit, the
802 maintenance of bedforms, and especially of bars, would be disconnected from the events' transport
803 distances or would result from successive minor changes. These findings do not support the hypothesis
804 of Pyrcce and Ashmore (2003a, 2005) claiming that, for full mobile events, bedload should be transported
805 on distances at least equal to the distance between two successive bars, equivalent to the length of a
806 riffle-pool unit, and that bedform maintenance depends on such events. More field and flume
807 experiments should now be undertaken to check our results and to better understand the suite of
808 conditions allowing the coexistence of an active, well-developed riffle-pool-bar morphology and low
809 bedload distances during full mobility events.

810 Further, we uncovered evidence for considerable decadal-scale variability in the bedload
811 velocity obtained from tracers according to the year of injection, the type of geomorphic unit that got
812 injected, or the method of injection (cluster or cross-section). This emphasises why it is imperative to
813 systematically consider the moment, location, and way tracers are seeded for the sound inference and
814 interpretation of bedload velocity. In addition to exemplifying the well-known strong spatiotemporal
815 variability of bedload processes, our results also highlight persistent effects upon velocity estimation
816 from where, when and how tracers are injected, especially for low-energy rivers.

817 Lastly, our PIT-tags survey provides new empirical data on the long-term dispersion of surface
818 seeded tracers at the decade temporal scale, this rarely documented in the literature, even less for low-
819 energy rivers (Bradley, 2017; Haschenburger, 2011a, 2013; McDowell and Hassan, 2020; Papangelakis
820 and Hassan, 2016). The right-skewed and heavy tailed velocity distribution reported here could indicate
821 a very incomplete mixing of tracers within the active layer. The vertical diffusion of tracers in such low-
822 energy rivers could therefore clearly be a more time-consuming process than previously reported for the
823 few rivers documented to date (Haschenburger, 2011b, 2012). Consequently, the long-term bedload
824 velocity estimation we gleaned from the PIT-tags data and bedload formulas should be viewed as
825 strongly overestimated. This pinpoints the difficulty in (1) obtaining robust information on the way of,
826 and velocity at which, the processes of advection-diffusion of the bedload occur in the long-term; and
827 (2) extrapolating to long-term velocities from the PIT-tags velocities recorded over just a few years.

828

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833

834 **References**

835 Andrews ED, Nankervis JM. 1995. Effective discharge and the design of channel maintenance flows
836 for gravel-bed rivers. In: Costa, J.E., Miller, A.J., Potter, K.W., Wilcock, P.R. (Eds.), *Natural and*
837 *Anthropogenic Influences in Fluvial Geomorphology*. American Geophysical Union, Washington
838 D.C., pp. 151-164.

839 Arnaud, F., Piégay, H., Vaudor, L., Bultingaire, L., Fantino, G., 2015. Technical specifications of low-
840 frequency radio identification bedload tracking from field experiments: differences in antennas,
841 tags and operators. *Geomorphology* 238, 37-46.

842 Ashmore, P.E., Church, M., 1998. Sediment transport and river morphology: A paradigm for study. In:
843 Klingeman, P.C., Beschta, R.L., Komar, P.D., Bradley, J.B. (Eds.), *Gravel-bed Rivers in the*
844 *Environment*. Water Resources Publications: Highlands Ranch, Colorado, pp. 115-148.

845 Baker, V.R., 1977. Stream-channel response to floods, with examples from central Texas. Geological
846 Society of American Bulletin 88 (8), 1057-1071.

847 Beechie, T.J., 2001. Empirical predictors of annual bed load travel distance, and implications for
848 salmonid habitat restoration and protection. Earth Surface Processes and Landforms 26, 1025-
849 1034.

850 Booker, D.J., Sear, D.A., Payne, A.J., 2001. Modelling three-dimensional flow structures and patterns
851 of boundary shear stress in a natural pool-riffle sequence. Earth Surface Processes and Landforms
852 26, 553-576.

853 Bradley, D.N., 2017. Direct observation of heavy-tailed storage times of bed load tracer particles causing
854 anomalous superdiffusion. Geophysical Research Letters 44, 12,227–12,235.

855 Bradley, D.N., Tucker, G.E., 2012. Measuring gravel transport and dispersion in a mountain river using
856 passive radio tracers. Earth Surface Processes and Landforms 37, 1034-1045.

857 Buffington, J.M., Lisle, T.E., Woodsmith, R.D, Hilton, S., 2002. Controls on the size and occurrence of
858 pools in coarse-grained forest rivers. River Research and Applications 18, 507-531.

859 Buffington, J.M., Montgomery, D.R., 1999. Effects of sediment supply on surface textures of gravel-
860 bed rivers. Water Resources Research 35 (11), 3523-3530.

861 Byrne, C.F., Pasternack, G.B., Guillon, H., Lane, B.A., Sandoval-Solis, S., 2021. Channel constriction
862 predicts pool-riffle velocity reversals across landscapes. Geophysical Research Letters, 48,
863 e2021GL094378.

864 Caamaño, D., Goodwin, P., Buffington, J.M., Liou, J.C.P., Daley-Laursen, S., 2009. Unifying criterion
865 for the velocity reversal hypothesis in gravel-bed rivers. Journal of Hydraulic Engineering 135 (1),
866 66-70.

867 Carling, P., 1991. An appraisal of the velocity-reversal hypothesis for stable pool-riffle sequences in the
868 River Severn, England. Earth Surface Processes and Landforms 16, 19-31.

869 Carling, P., Wood, N., 1994. Simulation of flow over pool-riffle topography: A consideration of the
870 velocity reversal hypothesis. Earth Surface Processes and Landforms 19, 319-332.

871 Carson, M.A., Griffith, G.A., 1989. Gravel transport in the braided Waimakariri River: Mechanisms,
872 measurements and predictions. Journal of hydrology 109, 201-220.

873 Chapuis, M., Bright, C.J., Hufnagel, J., MacVicar, B., 2014. Detection ranges and uncertainty of passive
874 Radio Frequency Identification (RFID) transponders for sediment tracking in gravel rivers and
875 coastal environments. *Earth Surface Processes and Landforms* 39 (15), 2109-2120.

876 Church, M., 2006. Bed material transport and the morphology of alluvial river channels. *Annual Review*
877 *of Earth and Planetary Sciences* 34, 325-354.

878 Church, M., Haschenburger, J.K., 2017. What is the “active layer”? *Water Resources Research* 53, 5–
879 10.

880 Church, M., Hassan, M.A., 2002. Mobility of bed material in Harris Creek. *Water Resources Research*
881 38 (11), 1237.

882 Church, M., Hassan, M.A., Wolcott, J.F., 1998. Stabilizing self-organized structures in gravel-bed
883 stream channels: field and experimental observations. *Water Resources Research* 34, 3169-3179.

884 Church, M., Mclean, D.G., Wolcott, J.F., 1987. River bed gravels: Sampling and analysis. In: Thorne,
885 C.R., Bathurst, J.C., Hey, R. (Eds.), *Sediment transport in gravel-bed rivers*. Wiley & Sons,
886 Chichester, pp. 43-79.

887 Clifford, N.J., 1993. Differential Bed Sedimentology and the Maintenance of Riffle-pool Sequences.
888 *Catena* 20, 447-468.

889 Clifford, N.J., Richards, K.S., 1992. The reversal hypothesis and the maintenance of pool-riffle
890 sequences: A review and field appraisal. In: Carling, P.A., Petts, G.E. (Eds.), *Lowland floodplain*
891 *ivers: Geomorphological perspectives*. Wiley & Sons, Chichester, pp. 43-70.

892 Curran, J.C., Wilcock, P.R., 2005. Effect of sand supply on transport rates in a gravel-bed channel.
893 *Journal of Hydraulic Engineering* 131 (11), 961-967.

894 Delignette-Muller, M.L., Dutang, C., 2015. fitdistrplus: An R Package for Fitting Distributions. *Journal*
895 *of Statistical Software* 64 (4), 1–34.

896 Dépret, T., 2014. Fonctionnement morphodynamique actuel et historique des méandres du Cher. Ph.
897 Thesis, Paris 1 – Panthéon-Sorbonne University.

898 Dépret, T., Gautier, E., Hooke, J., Grancher, D., Virmoux, C., Brunstein, D., 2015. Hydrological controls
899 on the morphogenesis of low-energy meanders (Cher River, France). *Journal of Hydrology* 531,
900 877–891.

901 Dépret, T., Gautier, E., Hooke, J., Grancher, D., Vermoux, C., Brunstein, D., 2017. Causes of planform
902 stability of a low-energy meandering gravel-bed river (Cher River, France). *Geomorphology* 285,
903 58–81.

904 Dietrich, W.E., Kirchner, J.W., Ikeda, H., Iseya, F., 1989. Sediment supply and the development of the
905 coarse surface layer in gravel-bedded rivers. *Nature* 340, 215-217.

906 Direction Départementale de l'Allier, 1998. Plan de Prévention du Risque Inondation du Val de Cher.

907 Downs, P.W., Soar, P.J., 2021. Beyond stationarity: influence of flow history and sediment supply on
908 coarse bedload transport. *Water Resources Research* 57(2), e2020WR027774.

909 Eaton, B.C., Lapointe, M.F., 2001. Effects of large floods on sediment transport and reach morphology
910 in the cobble-bed Sainte Marguerite River. *Geomorphology* 40, 291-309.

911 Einstein, H.A., 1937. Bed load transport as a probability problem. PhD Dissertation, Federal Institute of
912 Technology, Zurich.

913 Ferguson, R.I., 2003. The missing dimension: effects of lateral variation on 1-D calculations of fluvial
914 bedload transport. *Geomorphology* 56, 1–14.

915 Ferguson, R.I., Bloomer, D.J., Hoey, T.B., Werrity, A., 2002. Mobility of river tracer pebbles over
916 different timescales. *Water Resources Research* 38 (5), 1045.

917 Ferguson, R.I., Hoey, T.B., 2002. Long-term slowdown of river tracer pebbles: Generic models and
918 implications for interpreting short-term tracer studies. *Water Resources Research* 38 (8), 1142.

919 Frings, R.M., Vollmer, S., 2017. Guidelines for sampling bedload transport with minimum uncertainty.
920 *Sedimentology* 64, 1630–1645.

921 Gaeuman, D.A., Schmidt, J.C., Wilcock, P.R., 2003. Evaluation of in-channel gravel storage with
922 morphology-based gravel budgets developed from planimetric data. *Journal of Geophysical*
923 *research* 108 (F1), 6001.

924 Gilet, L., Gob., F., Gautier, E., Houbrechts, G., Vermoux, C., Thommeret, N., 2020. Hydro-
925 morphometric parameters controlling travel distance of pebbles and cobbles in three gravel bed
926 streams. *Geomorphology* 358, 107117.

927 Guillon, H., 2016. Origine et transport des sédiments dans un bassin versant alpin englacé (Glacier des
928 Bossons, France) : une quantification couplant mesures hydro-sédimentaires haute-résolution,

929 suivi radio-fréquence de galets, teneur en nucléides cosmogéniques et méthodes probabilistes. Ph.
930 Thesis, Grenoble Alpes University.

931 Haschenburger, J.K., 2011a. The rate of fluvial gravel dispersion. *Geophysical Research Letters* 38,
932 L24403.

933 Haschenburger, J.K., 2011b. Vertical mixing of gravel over a long flood series. *Earth Surface Processes*
934 and *Landforms* 36, 1044-1058.

935 Haschenbuger, J.K., 2013a. Tracing river gravels: Insights into dispersion from a long-term field
936 experiment. *Geomorphology* 200, 121-131.

937 Haschenbuger, J.K., 2017. Streambed Disturbance over a Long Flood Series. *River Research and*
938 *Applications* Volume 33 (5), 753-765.

939 Haschenburger, J.K., Wilcock, P.R., 2003. Partial transport in a natural gravel bed channel. *Water*
940 *Resources Research* 39, 1020.

941 Hassan, M.A., Bradley, D.N., 2017. Geomorphic controls on tracer particle dispersion in gravel bed
942 rivers. In: Tsutsumi, D., Laronne, J.B. (Eds.), *Gravel-Bed Rivers, Processes and Disasters*. Wiley-
943 Blackwell, UK, pp. 439-466.

944 Hassan, M.A., Church, M., 1992. The movement of individual grains on the streambed. In: Bili P, Hey,
945 R.D., Thorne, C.R., Tacconi, P. (Eds.), *Gravel-bed rivers III: Dynamics of gravel-bed rivers*. John
946 Wiley & Sons, Chichester, UK. pp. 159-175.

947 Hassan, M.A., Church, M., 2001. Sensitivity of bed load transport in Harris Creek : seasonal and spatial
948 variation over a cobble-gravel bar. *Water Resources Research* 37 (3), 813-825.

949 Hassan, M.A., Church, M., Schick, A.P., 1991. Distance of movement of coarse particles in gravel bed
950 streams. *Water Resources Research* 27 (4), 503-511.

951 Hassan, M.A., Saletti, M., Johnson, J.P.L., Ferrer-Boix, C., Venditti, J.G., Church, M., 2020.
952 Experimental insights into the threshold of motion in alluvial channels: Sediment supply and
953 streambed state. *Journal of Geophysical Research: Earth Surface* 125, e2020JF005736.

954 Hassan, M.A., Voepel, H., Schumer, R., Parker, G., Fraccarollo L., 2013. Displacement characteristics
955 of coarse fluvial bed sediment. *Journal of Geophysical Research Earth Surface*, 118 (1), 155-165.

- 956 Hassan, M.A., Woodsmith, R.D., 2004. Bed load transport in an obstruction-formed pool in a forest,
957 gravelbed stream. *Geomorphology* 58, 203-221.
- 958 Haynes, H., Pender, G., 2007. Stress history effects on graded bed stability. *Journal of Hydraulic*
959 *Engineering* 133 (4), 343-349.
- 960 Hill, B., 1975. A simple general approach to inference about the tail of a distribution. *The Annals of*
961 *Statistics* 3 (5), 1163–1174.
- 962 Hodge, R.A., Sear, D.A., Leyland, J., 2013. Spatial variations in surface sediment structure in riffle–
963 pool sequences: A preliminary test of the Differential Sediment Entrainment Hypothesis (DSEH).
964 *Earth Surface Processes and Landforms* 38 (5), 449-465.
- 965 Houbrechts, H., Levecq, Y., Peeters, A., Hallot, E., Van Campenhout, J., Denis, A.C., Petit, F., 2015.
966 Evaluation of long-term bedload virtual velocity in gravel-bed rivers (Ardenne, Belgium).
967 *Geomorphology* 251, 6-19.
- 968 Houbrechts, G., Levecq, Y., Petit, F., 2012. Characterisation of sand transport in gravel-bed rivers using
969 iron slag dated by historical studies. 2012 EGU General Assembly 2012, Vol. 14, EGU2012-4546.
- 970 Houbrechts, G., Levecq, Y., Vanderheyden, V., Petit, F., 2011. Long-term bedload mobility in gravel-
971 bed rivers using iron slag as a tracer. *Geomorphology* 126, 233-244.
- 972 Iwasaki, T., Yamaguchi, S., Yabe, H., 2018. Numerical simulation of bedload tracer transport associated
973 with sand bar formation, bank erosion, and channel migration. *River Flow 2018 - Ninth*
974 *International Conference on Fluvial Hydraulics E3S Web of Conferences* 40, 02013.
- 975 Jackson, W.L., Beschta, R.L., 1984. Influences of Increased Sand Delivery on the Morphology of Sand
976 and Gravel Channels. *JAWRA Journal of the American Water Resources Association* 20, 527–
977 533.
- 978 Kasprak, A., Wheaton, J.M., Ashmore, P.E., Hensleigh, J.W., Peirce, S., 2015. The relationship between
979 particle travel distance and channel morphology: Results from physical models of braided rivers.
980 *Journal of Geophysical Research: Earth Surface* 120, 55-74.
- 981 Keller, E.A., 1971. Areal sorting of bed-load material: The hypothesis of velocity reversal. *Geological*
982 *Society of America Bulletin* 82, 753-756.

- 983 Keller, E.A., Florsheim, J.L., 1993. Velocity-reversal hypothesis: A model approach. *Earth Surface*
984 *Processes and Landforms* 18,733-740.
- 985 Knighton, D., 1998. *Fluvial forms and processes, a new perspective*. Arnold, London, UK, 383 p.
- 986 Kochel, R.C., 1988. Geomorphic impact of large floods: review and new perspectives on magnitude and
987 frequency. In: Baker, V.R., Kochel, R.C., Patton, P.C. (Eds.), *Flood Geomorphology*. Wiley &
988 Sons, New York, pp. 169-187.
- 989 Kondolf, G.M., Mathews, W.V.G., 1986. Transport of tracer gravels on a coastal California river.
990 *Journal of Hydrology* 85, 265-280.
- 991 Konrad, C.P., Booth, D.B., Burges, S.J., Montgomery, D.R., 2002. Partial entrainment of gravel bars
992 during floods. *Water Resources Research* 38 (7), 1104.
- 993 Lajeunesse, E., Devauchelle, O., Houssais, M., Seizilles G., 2013. Tracer dispersion in bedload
994 transport. *Advances Geosciences* 37, 1–6.
- 995 Lajeunesse, E., Devauchelle, O., James, F., 2018. Advection and dispersion of bed load tracers. *Earth*
996 *Surface Dynamics* 6, 389–399.
- 997 Lamarre, H., Roy, A.G., 2008. The role of morphology on the displacement of particles in a step–pool
998 river system. *Geomorphology* 99, 270-279.
- 999 Lane, S.N., Richards, K.S., Chandler, J.H., 1995. Morphological estimation of the time-integrated bed
1000 load transport rate. *Water Resources Research* 31 (3), 761-772.
- 1001 Larue, J. P., 1981. Les nappes alluviales de la vallée du Cher dans le bassin de Montluçon. *Norois* 111,
1002 345–360.
- 1003 Larue, J. P., 2011. Longitudinal profiles and knickzones: the example of the rivers of the Cher basin in
1004 the northern French Massif Central. *Proceedings of the Geologists' Association* 122, 125–142.
- 1005 Leopold, L.B., 1992. Sediment size that determines channel morphology. In: Bili, P., Hey, R.D., Thorne,
1006 C.R., Tacconi, P. (Eds.), *Gravel-bed rivers III: Dynamics of gravel-bed rivers*. John Wiley & Sons,
1007 Chichester, UK, pp. 297-311.
- 1008 Lewin, J., 1989. Floods in fluvial geomorphology. In: Beven, K., Carling, P. (Eds.), *Floods:*
1009 *Hydrological, Sedimentological and Geomorphological Implications*. Wiley & Sons, Chichester,
1010 pp. 265-284.

- 1011 Lisle, T.E., 1979. A sorting mechanism for a riffle-pool sequence: Summary. Geological Society of
1012 America Bulletin 90, 616-617.
- 1013 Lisle, T.E., Hilton, S., 1992. The volume of fine sediment in pools: An index of sediments supply in
1014 gravel-beds streams. JAWRA Journal of the American Water Resources Association 28 (2), 371-
1015 383.
- 1016 Lisle, T.E., Hilton, S., 1999. Fine bed material in pools of natural gravel bed channels. Water Resources
1017 Research 35 (4), 1291-1304.
- 1018 Lisle, T.E., Iseya, F., Ikeda, H., 1993. Response of a channel with alternate bars to a decrease in supply
1019 of mixed-size bed load: a flume experiment. Water Resources Research 29 (11), 3623-3629.
- 1020 Lisle, T.E., Nelson, J.M., Pitlick, J., Madej, M.A., Barkett, B.L., 2000. Variability of bed mobility in
1021 natural , gravel-bed channels and adjustments to sediment load at local and reach scales model to
1022 examine relations between local boundary shear stress channels with widely differing annual
1023 sediment values channels. Water Resources Research 36 (12), 3743-3755.
- 1024 MacVicar, B.J., Roy, A.G., 2007a. Hydrodynamics of a forced riffle pool in a gravel bed river: 1. Mean
1025 velocity and turbulence intensity. Water Resources Research 43, W12401.
- 1026 MacVicar, B.J., Roy, A.G., 2007b. Hydrodynamics of a forced riffle pool in a gravel bed river: 2. Scale
1027 and structure of coherent turbulent events. Water Resources Research 43, W12402.
- 1028 MacVicar, B.J., Roy, A.G., 2011. Sediment mobility in a forced riffle-pool. Geomorphology 125, 445-
1029 456.
- 1030 MacWilliams, Jr. M.L., Wheaton, J.M., Pasternack, G.B., Street, R.L., Kitanidis, P.K., 2006. Flow
1031 convergence routing hypothesis for pool-riffle maintenance in alluvial rivers. Water Resources
1032 Research 42, W10427.
- 1033 Magilligan, F.J., 1992. Thresholds and the spatial variability of flood power during extreme floods.
1034 Geomorphology 5, 373-390.
- 1035 Mao, L., 2012. The effect of hydrographs on bed load transport and bed sediment spatial arrangement.
1036 Journal of Geophysical Research 117, F03024.

- 1037 Masteller, C.C., Finnegan, N.J., Turowski, J.M., Yager, E.M., & Rickenmann, D., 2019.
1038 History-dependent threshold for motion revealed by continuous bedload transport measurements
1039 in a steep mountain stream. *Geophysical Research Letters* 46, 2583–2591.
- 1040 May, C.L., Pryor, S., 2014. Initial motion and bedload transport distance determined by tracking in a
1041 large regulated river. *River Research and Application* 30 (4), 508-520.
- 1042 McDowell C., Hassan, M.A., 2020. The influence of channel morphology on bedload path lengths:
1043 Insights from a survival process model. *Earth Surface Processes and Landforms* 45 (12), 2982–
1044 2997.
- 1045 McDowell, C., Gaeuman, D., Hassan, M.A., 2021. Linkages between bedload displacements and
1046 topographic change. *Earth Surface Processes and Landforms* 46 (15), 3127-3142.
- 1047 McEwan, I.K., Habersack, H.M., Heald, J.G.C., 2001. Discrete particle modelling and active tracers:
1048 New techniques for studying sediment transport as Lagrangian phenomenon. In: *Gravel Bed*
1049 *Rivers* V., Mosley, P. (Eds). New Zealand Hydrological Society: Wellington, NZ, pp. 339-360.
- 1050 McQueen, R., Ashmore, P., Millard, T., Goeller, N., 2021. Bed particle displacements and
1051 morphological development in a wandering gravelbed river. *Water Resources Research*, 57,
1052 e2020WR027850.
- 1053 Milan, D.J., 2013. Sediment routing hypothesis for pool-riffle maintenance. *Earth Surface Processes*
1054 *and Landforms* 38 (14), 1623-1641.
- 1055 Milan, D.J., Heritage, G.L., Large, A.R.G., 2002. Tracer pebble entrainment and deposition loci:
1056 Influence of flow character and implications for riffle–pool maintenance. In: Jones, S.J., Frostick,
1057 L.E. (Eds.), *Sediment flux to basins: Causes, controls and consequences*. Geological Society of
1058 London Special Publications 191, pp. 133-148.
- 1059 Milan, D.J., Heritage, G.L., Large, A.R.G., Charlton, M.E., 2001. Stage dependent variability in tractive
1060 force distribution through a riffle–pool sequence. *Catena* 44, 85-109.
- 1061 Monteith, H., Pender, G., 2005. Flume investigations into the influence of shear stress history on a
1062 graded sediment bed. *Water Resources Research* 41, W12401.
- 1063 Neill, C.R., 1971. River bed transport related to meander migration rates. *Journal of the Waterways,*
1064 *Harbors, and Coastal Engineering Division, ASCE* 97, 783-786.

1065 Neill, C.R., 1987. Sediment balance considerations linking long-term transport and channel processes.
1066 In: Thorne, C.R., Bathurst, J.C., Hey, R.D. (Eds.), Gravel-bed rivers III: Sediment transport in
1067 gravel-bed rivers. John Wiley & Sons, Chichester, UK, pp. 225-240.

1068 Nelson, P.A., Venditti, J.G., Dietrich, W.E., Kirchner, J.W., Ikeda, H., Iseya, F., Sklar, L.S., 2009.
1069 Response of bed surface patchiness to reductions in sediment supply. *Journal of Geophysical*
1070 *Research* 114, F02005.

1071 Ockelford, A.M., Haynes, H., 2013. The impact of stress history on bed structure. *Earth Surface*
1072 *Processes and Landforms* 38 (7), 717-727.

1073 Paintal, A.S., 1971. A stochastic model of bed load transport. *Journal of Hydraulic Research* 9 (4): 527-
1074 554.

1075 Papangelakis, E., Hassan M.A., 2016. The role of channel morphology on the mobility and dispersion
1076 of bed sediment in a small gravel-bed stream. *Earth Surface Processes and Landforms* 41 (15),
1077 2191-2206.

1078 Paphitis, D., Collins, M.B., 2005. Sand grain threshold, in relation to bed “stress history”: an
1079 experimental study. *Sedimentology* 52, 827-838.

1080 Parker, G., Klingeman, P.C., McLean, D.L., 1982. Bedload and size distribution in paved gravel-bed
1081 streams. *Journal of Hydraulic Engineering* 108, 544-571.

1082 Pelosi, A., Schumer, R., Parker, G., Ferguson, R.I., 2016. The cause of advective slowdown of tracer
1083 pebbles in rivers: Implementation of Exner-Based Master Equation for coevolving streamwise and
1084 vertical dispersion, *Journal of Geophysical Research* 121 (3), 623-637.

1085 Petit, F., 1987. The relationship between shear stress and the shaping of the bed of a pebble-loaded
1086 river, La Rulles – Ardenne. *Catena* 14, 453-468.

1087 Pierce, J.K., Hassan, M.A., 2020. Back to Einstein: Burial-induced three-range diffusion in fluvial
1088 sediment transport. *Geophysical Research Letters* 47, e2020GL087440.

1089 Pyrcce, R.S., Ashmore, P.E., 2003a. Particle path length distributions in meandering gravel-bed streams:
1090 Results from physical models. *Earth Surface Processes and Landforms* 28, 951-966.

1091 Pyrcce, R.S., Ashmore, P.E., 2003b. The relation between particle path length distributions and channel
1092 morphology in gravel-bed streams: A synthesis. *Geomorphology* 56, 167-187.

- 1093 Pyrcce, R.S., Ashmore, P.E., 2005. Bedload path length and point bar development in gravel-bed river
1094 models. *Sedimentology* 52, 839-857.
- 1095 Recking, A., 2010. A comparison between flume and field bed load transport data and consequences for
1096 surface-based bed load transport prediction. *Water Resources Research* 46, W03518.
- 1097 Recking, A., Piton, G., Vázquez-Tarrío, D., Parker, A., 2016. Quantifying the Morphological Print of
1098 Bedload Transport. *Earth Surface Processes and Landforms* 41 (6), 809-822.
- 1099 Reid, I., Frostick, L.E., Layman, J.T., 1985. The incidence and nature of bedload transport during flood
1100 flows in coarse-grained alluvial channels. *Earth Surface Processes and Landforms* 10, 33-44.
- 1101 Reid, S.C., Lane, S.N., Berney, J.M., Holden, J., 2007. The timing and magnitude of coarse sediment
1102 transport events within an upland, temperate gravel-bed river. *Geomorphology* 83, 152-182.
- 1103 Rickenmann, D., Recking, A., 2011. Evaluation of flow resistance in gravel-bed rivers through a large
1104 field data set. *Water Resources Research* 47 (7), W07538.
- 1105 Roseberry, J.C., Schmeeckle, M.W., Furbish, D.J., 2012. A probabilistic description of the bed load
1106 sediment flux: 2. Particle activity and motions. *Journal of Geophysical Research* 117, F03032.
- 1107 Schmidt, K.H., Gintz, D., 1995. Results of bedload tracer experiments in a mountain river. In: Hickin,
1108 E.J. (Eds.). *River Geomorphology*. John Wiley & Sons, Chichester, UK, pp. 37-54.
- 1109 Sear, D.A., 1996. Sediment transport process in pool-riffle sequences. *Earth Surface Processes and*
1110 *Landforms* 21, 241-262.
- 1111 Simon-Coinçon, R., Thiry, M., Quesnel, F., 2000. Paléopaysages et paléoenvironnements
1112 sidérolithiques du Nord du Massif central (France). *Comptes Rendus de l'Académie des Sciences*
1113 - Series IIA - Earth and Planetary Science 330, 693-700.
- 1114 Thompson D.M., 2007. The characteristics of turbulence in a shear zone downstream of a channel
1115 constriction in a coarse-grained forced pool. *Geomorphology* 83, 199-214.
- 1116 Thompson, D.M., 2011. The velocity-reversal hypothesis revisited. *Progress in Physical Geography* 35
1117 (1), 123-132.
- 1118 Thompson, D.M., MacVicar, B.J., 2020. Pool-Riffle. In: *Reference Module in Earth Systems and*
1119 *Environmental Sciences*, Elsevier.

- 1120 Thompson, D.M., Whol, E.E., 2009. The linkage between velocity patterns and sediment entrainment
1121 in a forced-pool and riffle unit. *Earth Surface Processes and Landforms* 34, 177-192.
- 1122 Thompson, D.M., Wohl, E.E., Jarrett, R.D., 1999. Velocity reversals and sediment sorting in pools and
1123 riffles controlled by channel constrictions. *Geomorphology* 27, 229-241.
- 1124 Turland, M., Cojean, R., Brulhet, J., Morice, E., Grolier, J., Lacour, A., 1989a. Carte géol. France
1125 (1/50000), feuille Hérisson (596), BRGM, Orléans.
- 1126 Turland, M., Hottin, A.M., Cojean, R., Ducreux, J.L., Debégli, N., d'Arcy, D., Mathis, V., Carroué,
1127 J.P., Piboule, M., 1989b. Notice explicative, Carte géol. France (1/50000), feuille Hérisson (596),
1128 BRGM, Orléans, 118 p.
- 1129 Vahidi, E., Rodríguez, J.F., Bayne, E., Saco, P.M., 2020. Oneflood is not enough: Pool-riffle self-
1130 maintenance under time- varying flows and non equilibrium multifractional sediment
1131 transport. *Water Resources Research* 56, e2019WR026818.
- 1132 Vayssière, A., Castanet, C., Gautier, E., Vermoux, C., Dépret, T., Gandouin, E., Develle-Vicent A.L.,
1133 Mokadem, F., Saulnier-Copard, S., Sabatier, P., Carcaud, N., 2020. Long-term readjustments of a
1134 sinuous river during the second half of the Holocene in northwestern Europe (Cher River, France).
1135 *Geomorphology* 370, 107395.
- 1136 Vázquez-Tarrío, D., Batalla, R.J., 2019. Assessing Controls on the Displacement of Tracers in Gravel-
1137 Bed Rivers. *Water* 11 (8), 1598.
- 1138 Vázquez-Tarrío, D., Piqué, G., Vericat, D., Batalla, R.J., 2021. The active layer in gravel-bed rivers: An
1139 empirical appraisal. *Earth Surface Processes and Landforms* 46 (2), 323-343.
- 1140 Vázquez-Tarrío, D., Recking, A., Liébault, F., Tal, M., Menéndez-Duarte, R., 2019. Particle transport
1141 in gravel-bed rivers: Revisiting passive tracer data. *Earth Surface Processes and Landforms* 44 (1),
1142 112-128.
- 1143 Whiting, P.J., Stamm, J.F., Moog, D.B., Orndorff, R.L., 1999. Sediment-transporting flows in headwater
1144 streams. *Geological Society of America Bulletin* 111 (3), 450-466.
- 1145 Wilcock, P.R., 1998. Two-fraction model of initial sediment motion in gravel-bed rivers. *Science* 280
1146 (5362), 410-412.

- 1147 Wilcock, P.R., Crowe, J.C., 2003. Surface-based transport model for mixed-size sediment. *Journal of*
1148 *Hydraulic Engineering* 129 (2), 120-128.
- 1149 Wilcock, P.R., McArdeell, B.W., 1993. Surface-based fractional transport rates: Mobilization thresholds
1150 and partial transport of a sand-gravel sediment. *Water Resources Research* 29 (4), 1297-1312.
- 1151 Wilkinson, S.N., Keller, R.J., Rutherford, I.D., 2004. Phase-shifts in shear stress as an explanation for
1152 the maintenance of pool-riffle sequences. *Earth Surface Processes and Landforms* 29, 737-753.
- 1153 Wolman, M.G., 1954. A method of sampling coarse river-bed material. *Transactions of the American*
1154 *Geophysical Union* 35 (6), 951-956.
- 1155 Wu, Z., Singh, A., Fu, X., Wang, G., 2019. Transient anomalous diffusion and advective slowdown of
1156 bedload tracers by particle burial and exhumation. *Water Resources Research* 55 7964–7982.
- 1157 Zhang, Y., Meerschaert, M.M., Packman, A.I., 2012. Linking fluvial bed sediment transport across
1158 scales. *Geophysical Research Letters* 39 (20), L20404.