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Controls Over Particle Motion and Resting Times of Coarse Bed Load Transport in a Glacier-Fed Mountain Stream

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1	Controls over particle motion and resting times of coarse bed load transport in a
2	glacier-fed mountain stream
3	
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18	
19	Abstract
20	Coarse bed load transport is a crucial process in river morphodynamics but is difficult to
21	monitor in mountain streams. Here we present a new sediment transport dataset obtained
22	from two years of field-based monitoring (2014-2015) at the Estero Morales, a high-
23	gradient stream in the central Chilean Andes. This stream features step-pool bed geometry and
24	a glacier-fed hydrologic regime characterized by abrupt daily fluctuations in discharge. Bed
25	load was monitored directly using Bunte samplers and by surveying the mobility of PIT
26	(passive integrated transponder) tags. We used the competence method to quantify the effective
27	slope, which is the fraction of the topographical slope responsible for bed load transport. This
28	accounts for only 10% of the topographical slope, confirming that most of the energy is
29	dissipated on macroroughness elements. We used the displacement lengths of PIT tags to
30	analyze displacement lengths and virtual velocity of a wide range of tracer sizes (38-415 mm).
31	Bed load transport in the Estero Morales show to be size-selective and the distance between

32 steps influences the displacement lengths of PIT tags. Displacement lengths were also used to

33 derive the statistics of flight distances and resting times. Our results show that the average

34 length of flight scales inversely to grain size. This contradicts Einstein's conjecture about the

35 linear relationship between grain size and intervals between resting periods in a steep step-pool

- 36 stream in ordinary flood conditions.
- 37

38 **1. Introduction**

Sediment transport, flow resistance, morphological structure and bed texture in rivers are interrelated in complex ways (Church, 2006). Coarse bed load transport in mountain streams is important for civil engineers, geomorphologists, river ecologists and managers, as it determines channel morphology and dynamics, while representing a significant component of flood-related costs (e.g. Badoux et al., 2014). However, monitoring bed load in the field can be logistically onerous, costly, and affected by a number of uncertainties (Dell'Agnese et al., 2014; Schneider et al., 2016; Vericat et

46 al., 2006; Magirl et al., 2015).

47 Predicting bed load is a difficult task, especially in steep mountain streams, where

48 channel beds are often organized in boulder-cascade and step-pool morphologies

49 (Brardinoni & Hassan, 2007; Buffington & Montgomery, 1997; Comiti & Mao, 2012).

50 Modelling the flow over such bed structures is still very challenging (Saletti et al.,

51 2016; Escauriaza et al., 2017), especially when levels of submergence are low (Comiti

et al., 2009; Wilcox et al., 2011). Modeling flow and bedload transport in mountain

53 stream is complex because patches of mobile sediments coexist with immobile

54 structures such as steps or isolated boulders (Laronne et al., 2001; Green et al., 2015;

55 Yager et al., 2012). Furthermore, critical dimensionless shear stress in mountain

streams (i.e. Shields stress) is very difficult to model, due to its high variability in time

57 (Masteller et al., 2019) and space (Monsalve et al., 2016), modulated by changes in

58 local channel slope (Lamb et al., 2008) and relative submergence of sediments (Lenzi

59 et al., 2006).

60 Only part of the flow power in mountain streams is available to entrain and transport

61 sediments (Rickenmann & Recking, 2011; Hohermuth & Weitbrecht, 2018), as some

62 energy is dissipated in various ways, including local hydraulic plunging and jumps

over steps (Comiti et al., 2009; Green et al., 2013; Monsalve et al., 2017). In such 63 complex hydraulic conditions, well known and established flow resistance approaches 64 65 and empirical bed load formulas are unlikely to be reliable (e.g. Yager et al., 2018). Considering that mountain streams are typically supply limited (Lenzi et al., 2004; 66 Turowski et al., 2009), bed load formulas fail to predict actual transport rates by orders 67 68 of magnitude in step-pool streams (Nitsche et al., 2011), since not all the energy 69 available is effective in transporting sediments. Because macroroughness can represent up to 90 % of the total flow resistance in step-pool streams (Church & Zimmermann, 70 2007), accounting for the portion of energy slope or flow resistance associated with 71 macroroughness can substantially increase the predictive power of empirical bed load 72 formulas (Chiari & Rickenmann, 2011). Using a large empirical database, 73 Rickenmann & Recking (2011) adopted Ferguson's approach (2007) for partitioning 74 flow resistance and splitting the friction factor into a base level value associated with 75 small-scale grain size roughness, and total value, which includes macroroughness, 76 such as step-pools. A similar breakdown can be applied to the topographical bed slope 77 (i_b) , from which an effective slope value (i_e) representing that part of the slope that 78 plays a role in coarse grain entrainment and movement can be extracted from the 79 remaining part of the slope $(i_b - i_e)$ associated with the mechanical power dissipated by 80 81 flow over immobile bed forms (Nitsche et al, 2011). Controls on sediment transport processes and dynamics in mountain streams can be 82 explored using tracers. Displacement lengths of tracers depend on particle size and 83 shear stress acting on them (e.g. Church & Hassan, 1992), but also on dimensionless 84 impulse (e.g. Phillips et al., 2013) and antecedent flow conditions (e.g. Mao et al., 85 86 2017). Displacement length depends also on channel morphology and especially on the spacing between bedforms (Vázquez-Tarrío et a., 2019). Displacement length is 87 88 needed to calculate the virtual velocity of sediment, that can be used to assess coarse bed material transport (e.g. Brenna et al., 2019). Velocity is called virtual because it is 89 90 usually calculated considering the interval between two consecutive surveys of tracers. However, tracers would naturally move through sequences of jumps and rest periods 91 92 (Einstein, 1950) that are observable in laboratory experiments (e.g. Fraccarollo & Hassan, 2019) but difficult to monitor in the field, with only a few exceptions (e.g. 93

94 Habersack, 2001; Maniatis et al., 2017).

In this paper, we use a new dataset of direct bed load surveys in a step-pool stream to 95 96 calculate the portion of slope effective in transporting sediments. We then use this slope value to analyze sediment transport dynamics, and apply the Einstein (1950) 97 and virtual velocity (Haschenburger & Church, 1998) approaches to characterize bed 98 load mobility. Both approaches are based on a Lagrangian perspective of sediment 99 100 transport which considers the motion of individual particles, and views moving grains as being almost independent of each other. While sliding, rolling and saltating over the 101 102 bed, sediments may also experience long resting periods on the surface, or may be temporarily buried as a result of temporal and spatial changes in the bed surface 103 104 profile (Singh et al., 2009; Wong et al., 2007). Both the Einstein and the virtual velocity formulations allow for predicting bed load transport rates and can be expressed 105 106 as a function of Shields stress, such as the MPM empirical bed load formula (Meyer-Peter & Muller, 1948; Wong & Parker, 2006). Here we will use field data to infer the 107 108 predictors of Einstein's statistical approach. The study was conducted at the Estero Morales, a small glacier-fed Andean step-pool stream in central Chile. Coarse 109 sediment mobility and transport rates were monitored in 2014-2015 using Bunte traps 110 and Passive Integrated Transponders (PIT) tagged clasts. The specific aims of the 111 paper are to i) use the competence approach to assess energy dissipation in step-pool 112 morphology by calculating the effective slope of the stream; ii) use the effective slope 113 to assess the increase in the predictive power of a widely applied formula in a system 114 with nearly unlimited sediment supply; iii) use the virtual velocity of PIT tracers to 115 assess sediment mobility; and iv) assess the resting times of PIT tags using both 116 117 particles that moved during the survey period and those that did not.

118

119 2. Study site and monitoring methods

120 *2.1. Study site*

121 The study was conducted in a reach of the Estero Morales, a high-gradient stream in

the central Chilean Andes (Figure 1). The basin extends for 27 km^2 , with an

123 elevation between 1780 and 4497 m.a.s.l. There are several relatively small glaciers

above 2700 m.a.s.l. (Mao & Carrillo, 2017). Mean annual rainfall, as measured in the

lower part of the basin, is around 550 mm. The climate is Mediterranean, with 125 precipitation occurring mainly as snowfall from April to September, with occasional 126 127 convective summer storms and late summer rain-on-snow events. The catchment is covered by snow for approximately 5 months a year, so discharge mainly results from 128 snowmelt in late spring (from late September to November) and glacier melt from 129 December to March. Bedrock geology of the catchment consists of volcanic 130 131 metamorphic rock and conglomerate-sand deposits. The stream in the study reach is steep (0.14 m/m) and the channel width is approximately 7 m. The study reach is 132 760 m long and is located just upstream from the confluence with the El Volcan 133 River (Figure 1). The stream flows through coarse lag deposits and features typical 134 135 of step-pool sequences (Figure 2). In 2015, we counted 72 step-pool units over the 760 m-long reach, with an average spacing of approximately 10.5 m. The step-pool 136 sequences are stable during ordinary glacial melting floods and are only prone to 137 dislodgement and rearrangement during extreme late summer events. The site was 138 139 monitored continuously from July 2013 until April 2016, when a high-magnitude event caused considerable channel changes and destroyed the monitoring devices. 140

141

142 2.2. Bed load monitoring and sampling using Bunte traps

The study reach was equipped with a pressure transducer and a multiparameter water 143 quality probe (OTT Hydrolab MS5) to monitor water temperature, electrical 144 conductivity, and turbidity with a 10-min time resolution (see Mao & Carrillo, 2017). 145 A 0.5-m-long acoustic pipe sensor was also installed (Mao et al., 2016). A series of 146 direct bed load samples were collected near the monitoring cross-section using 0.3 m-147 wide, 0.25 m-high Bunte-type traps (Bunte et al., 2004). Two traps were located at 148 points near the bed load acoustic pipe sensor (Figure 2), located on the sampling site 149 (Figure 1). Samples were collected during the glacier melting season (January to 150 March) in 2014 and 2015 at discharges ranging from 1 to 3 m³ s⁻¹. The highest 151 discharge monitored was roughly the bankfull discharge. The highest discharge in the 152 Estero Morales (approx. $7 \text{ m}^3 \text{ s}^{-1}$) was recorded in April 2016 and was able to destroy 153 and recreate step-pools (Mao & Carrillo, 2017). Samplings was concentrated in a few 154 days to monitor bed load at a wide range of transport rates. Multiple samples were 155

taken, especially during the rising limb of hydrographs. The duration of bed load
sampling ranged from 1 to 120 minutes depending on the discharge. The amount of
trapped sediment ranged from 0.1 to 49.7 kg. The sediment collected in the traps was
dried, weighed and later sieved in the laboratory.

160

161 2.3. Bed load monitoring using tracers

162 Coarse sediment mobility in Estero Morales was also investigated using natural clasts equipped with 23-mm-long radio frequency identification (RFID) passive integrated 163 164 transponder tags (Hassan et al., 2013; Liebault et al., 2012). PIT tags are transmitters without batteries that emit an identification codes by radio frequencies, which are 165 166 detected and recorded. The RFID PIT system uses electromagnetic fields to automatically identify tags. A commercially available mobile RFID system 167 168 consisting of a reader unit, a battery suppling the system, a screen, and a short-range antenna (0.5 m in diameter), was used to determine the position of PIT tags (Figure 169 170 2). The antenna can generally recognize tracers in a range of 0.3 m, although this varies depending on the position of the clast relative to the antenna, the battery charge, 171 and interference among PIT tags (see Chapuis et al., 2014). 172

PIT tags were inserted in holes drilled in the clasts and closed using epoxy glue. The 173 three main dimensions of the clasts (a - long, b - medium, c - short) were measured 174 with a caliper and the clasts were weighed to determine the relationship between b-175 axis diameter and weight. Only sediments collected in the Estero Morales were used 176 to produce PIT tracers for this study. The tracers were also spray-painted to enhance 177 their visibility in the channel bed. The PIT tags were placed in the upstream end of 178 the study reach and ended up at the confluence with the El Volcan River. As in other 179 studies (e.g. Dell'Agnese et al., 2015), the PIT tags were laid in the channel bed and 180 gently pushed by foot into the surface sediment material to prevent them from 181 protruding abnormally from the bed. PIT tags were always inserted at three transects 182 183 (spaced 0.5 m apart) at the upstream end of the reach. The operator equipped with 184 the mobile antenna scanned the channel bed surface for tags. When a tag was 185 detected, vertical and horizontal distance were measured from wooden stakes with a 186 laser rangefinder. Overall, 32 wooden stakes were located along the study site, and

187 all were georeferenced using DGPS. At the beginning of the measuring periods in January 2014 and January 2015, 461 and 395 PITs respectively were placed in the 188 189 study reach during low-flow conditions. PIT positions were surveyed 20 times in 2014 and 13 times in 2015. Each survey took one to two days, and required intense in-190 channel surveys, carried out by two operators, who worked until 2 PM, after which 191 increased discharge made the search for PITs unsafe. The flow regime during the 192 193 monitoring search for PITs was always characterized by daily fluctuations in discharge. The surveys always started from the upstream end of the monitored reach. 194 195 Due to the rapid increase in water discharge in the early afternoon, it was generally possible to survey only the upper half of the reach. On one occasion, it was possible to 196 197 survey the entire reach. Figure 3 shows the size distribution of the tracers, and the bed surface and subsurface sediments. A surface grid-by-number sampling revealed that 198 the sediments are coarse and poorly sorted ($D_{16} = 20$ mm; $D_{50} = 59$ mm; $D_{84} = 318$ 199 mm; $D_{90} = 448$ mm; Figure 3). A volume-by-weight subsurface sampling showed that 200 201 bed material grain size distribution is finer than that of the surface, demonstrating a certain rate of surface armoring ($D_{50-sub} = 25$ mm; Figure 3). The b-axes of tracers 202 range from 27 to 420 mm. The lower value of this range was determined by the length 203 of the PIT tags inserted in the clasts (23 mm), and the upper value by the size of 204 pebbles in which PITs could be inserted and then transported in the study site. 205

206

207 3. Conceptual and theoretical framework: unifying the statistical motion and 208 virtual velocity approaches

In this section, we introduce Einstein's (1950) bed load formula, and present the way 209 field data collected with the PIT tracers was used to back-calculate the predictors that 210 appear in the formula. Einstein's pioneering approach describes the bed load transport 211 rate as a function of two predictors that consider the probabilistic distribution of mean 212 particle displacements \overline{L} and $\overline{T^{-1}}$, being T the resting time. L represents the distance 213 that a particle runs without pausing during its movement from an entrainment to the 214 subsequent deposition. We refer to this as flight distance, whereas in other studies (e.g. 215 216 Bradley et al., 2010; Habersack, 2001; Nikora et al., 2002) it is termed step or travel distance, quick length step, intermediate trajectories, and travel distance. T is the 217

resting time, i.e. the time interval from a first deposition to the next entrainment,

- whereas T^{1} represents the probability for a single grain at rest to be entrained in a unit
- of time. The Einstein expression has two further factors, which are constant for a given
- bed, i.e. the number N of grains paving a unit area and the reference volume V of a
- single particle. N can be evaluated as the ratio between bed particle concentration c_b
- 223 (volumetric percentage of solid volumes) and the particle reference cross section A.
- Following the above notations, the Einstein (1950) formulation for the unit volumetric transport rate q_s (in m²/s) is expressed as:
 - $q_s = \overline{T^{-1}} \forall \frac{c_b}{A} \overline{L} \tag{1}$

In order to make Eq. 1 applicable to all i^{th} grain size classes of a poorly sorted mixture, we add a factor f_i , which indicates the volumetric frequency of the class size D_i , resulting in:

226

$$q_{si} = \overline{T_i^{-1}} \forall_i \frac{c_b}{A_i} \overline{L_i} f_i \tag{2}$$

where V_i is the reference volume of a single particle in the i^{th} grain size class. The term $\overline{T_i}^{-1} \frac{c_b}{A_i}$ can be termed E_i , which represents the number of particles entrained per unit of time and unit of bed-surface area. Unless the probability distribution of the resting times is uniform, it holds that:

235

$$\overline{T_{l}^{-1}} \neq \frac{1}{\overline{T_{l}}} \tag{3}$$

The ratio \forall_i / A_i is equal to αD_i , α being a constant. Indeed, for nearly spherical-shaped grains it holds that $\forall_i / A_i = \frac{4}{3}\pi \frac{D_i^3}{8} / \pi \frac{D_i^2}{4} = \frac{2D_i}{3}$. Thus, it is generally true that α is close to unity and depends on the shape of the particle. Eventually, Eq. 2 becomes:

 $q_{si} = \alpha D_i c_b \overline{T_i^{-1}} f_i \overline{L_i}$ (4)

The distribution of T_i can be obtained from the dataset collected in the field using the tracers (see Section 6).

The unit volumetric transport rate q_s (per unit of width of the active channel) in the virtual velocity approach can be calculated for well sorted sediments as follows (Haschenburger & Church, 1998; Liebault & Laronne, 2008; Hassan & Bradley,

245 2017):

$$q_s = V_v \delta c_b \tag{5}$$

where c_b is the volumetric concentration of the sediments in the bed and δ is the 247 thickness of the active layer of sediments. The virtual velocity of bed load V_{v} is 248 249 calculated as the ratio between the total displacement ℓ travelled by a particle in a certain time interval t, and t itself. Because particles are entrained, move and rest 250 several times during a transport event, it holds that t is much longer than the above 251 defined T and that the total displacement ℓ is much longer than the previously 252 defined \overline{L} . Virtual velocity V_{ν} is thus the averaged velocity, over both motion and 253 resting periods, of a large ensemble of grains running their trajectory. The transport 254 of each ith grain size fraction of a poorly sorted bed mixture, given its volumetric 255fraction f_i in the bed mixture, and its representative grain size D_i , is given by: 256

 $q_{si} = V_{vi} \delta c_b f_i \tag{6}$

In this paper, we use the data collected with PIT tracers in sequential surveys to calculate the virtual sediment velocity V_{ν} , calculating T_T as the time elapsed from when a trace was inserted in the bed (or the previous survey) until the last time it was detected, and l_T as the distance traveled from the release (or the previous survey) to the position at which it was last recovered (as is usually done in literature, see

263 Vasquez-Tarrio et al., 2019). From Eq. (6) it can be drawn that:

264

$$V_{vi} = q_{si}/c_b \delta f_i \tag{7}$$

With the aim of establishing a relation between the two approaches, Eqs. 2 and 7 canbe combined as:

$$q_{si} = \alpha D_i c_b \overline{T_i^{-1}} f_i \overline{L_i} = V_{vi} \delta c_b f_i$$

269

267

$$V_{\nu i} \approx \overline{T_i^{-1}} \, \overline{L_i} \, \frac{\alpha D_i}{\delta} \tag{9}$$

(8)

which displays a straightforward interpretation of virtual velocity in terms of
predictors in Einstein's approach. Interestingly, Eq. 9 applies to both well and poorly
sorted bed mixtures.

273

4. Effective slope and predictive power of empirical bed load formulas

Around two thirds of the total energy in steep streams can be dissipated in the

hydraulics generated by step-pool sequences (e.g. Wilcox et al., 2011), and this

dissipated energy does not contribute to generating bed load transport. Here we apply

a slope decomposition approach (as in Nitsche et al., 2011), in which the mean topographic longitudinal slope i_b is split into the effective slope involved in sediment

- transport (i_e) and the component of slope that is absorbed in form and spill drags (i_f) :
- 281

$$i_b = i_e + i_f \tag{10}$$

To calculate the effective slope i_e , we used the competence approach for incipient 282 sediment motion, which considers the largest particle transported and collected during 283 284 bed load samplings conducted under different hydrometric conditions (Andrews, 1983; 285 Batalla & Martin-Vide, 2001; Mao et al., 2008). In the case of the Estero Morales, we 286 used the size of the coarsest grain captured with the Bunte traps (D_{100}) and the shear stress acting on the bed calculated using the depth-slope approach (Wilcock, 1993), 287 which involves the use of the water depth at the time of sampling (h), topographical 288 289 slope (i_b) , the acceleration constant due to gravity (g) and water density (ρ) . The shear stress associated with sediment transport is calculated as: 290

291

$$\tau = h i_h g \rho \tag{11}$$

Here we adopted the flow competence method, which assumes that the local flow conditions at the moment of bedload sampling can be considered critical for the entrainment and transport of the coarsest sediment trapped in the sampler (Andrews, 1983; Lenzi et al., 2006). In other words, the shear stress during a bedload sampling is critical for the coarsest particles found in the Bunte trap (D_{100}), and thus the critical Shields stress (θ_{cr}) can be calculated as:

298

301

 $\theta_{cr} = \frac{h \cdot i_b}{\Delta \cdot D_{100}} \tag{12}$

where $\Delta = \rho_s / \rho - 1$ and ρ_s is the density of sediments. If i_e , rather than i_b , is considered in Eq. 12, the effective slope can be calculated as:

- $i_e = \frac{\theta_{cr} \cdot \Delta \cdot D_{100}}{h} \tag{13}$
- which shows that the ratio D_{100}/h is proportional to i_e , provided that θ_{cr} is assumed to be constant. Figure 4 shows the relationship between transported D_{100} and the water depth at the time of direct bed load sampling conducted in 2014 and 2015. The size of the coarsest transported sediments increases with the flow depth, although the predictive power of a linear regression is very weak ($R^2 = 0.29$). The considerable scatter of data is consistent with previous observations (e.g., Mao et al., 2008). On the one hand, this could be attributed to the reduced trapping efficiency of Bunte

samplers, especially at higher bed load transport rates due to shorter sampling times 309 and the fact that D_{100} approaches the size of the sampler intake. On the other hand, the 310 311 scatter observable in Figure 4 is also due to the natural grain size variability of upstream sediment delivery sources, dynamics of destruction of sediment clusters, 312 changes in sediment imbrication, and temporal variability of local shear stress. As 313 well, the flow conditions at the bed load sampling point are unlikely to represent the 314 315 local conditions well in the place where the sediments were entrained, causing further scatter in the relationship between the coarsest clast captured by the sampling basket 316 317 and the water stage at the site of monitoring.

318 The effective slope can be calculated by assuming a certain value of the critical

319 Shields parameter θ_{cr} unaffected by grain sorting, particle size interactions, channel

320 slope or high relative roughness. Although the critical Shields parameter can vary

321 locally, ranging from 0.03 to 0.09 (Buffington & Montgomery, 1997), if an

intermediate value of 0.05 is assumed, Eq. 13 provides a calculated value of i_e of 0.018

323 m/m. Interestingly, this value of effective slope is almost one order of magnitude

lower than the actual mean topographic channel slope ($i_b = 0.14$ m/m), revealing that

325 only a small part of the available bed shear stress is actually associated with sediment

326 transport during ordinary daily floods in the Estero Morales. This is well in agreement

327 with the high level of energy dissipation associated with bedforms (Chin & Wohl,

328 2005; Comiti et al., 2009), which can account for 80 % of the total flow resistance

329 (e.g. Canovaro et al., 2007; Chiari et al., 2010; Wilcox et al., 2011).

330 Our calculation of effective slope can be compared with the effective slope estimated

using the power-function relation proposed by Rickenmann and Recking (2011):

332

$$\frac{i_e}{i_b} = (i_b^{-z}p)^e \tag{14}$$

where p = 0.07, z = 0.47 and e = 1.2 are calibrated using a large empirical dataset (Chiari & Rickenmann, 2011; Rickenmann & Recking, 2011). For the Estero Morales, Eqs. 13 and 14 provide an estimates of i_e/i_b , of around 0.13, remarkably similar to the ratio resulting from the use of the value of i_e (0.018 m/m, see above), confirming that the analysis based on competence is sound. Notably, Nitsche et al. (2011) combined several bed load transport and flow resistance equations to account for flow energy dissipation of step-pool morphology in steep streams. They recommended using the 340 slope decomposition approach (i.e., Rickenmann & Recking 2011) to increase the

341 predictive power of bed load formulas that generally overestimate bed load transport

in mountain streams by orders of magnitude (e.g. Vázquez-Tarrío & Menéndez-

343 Duarte, 2015). Here we use effective slope to assess the increase in the predictive

power of a widely-applied formula in a system with nearly unlimited sediment supply,

and to derive the reference values needed for the later development of the present

work. Specifically, we applied the Meyer-Peter & Muller equation (Meyer-Peter & Muller, 1948; Wong & Parker, 2006, hereafter MPM) using both the effective slope i_e and the topographical slope i_b , to calculate the value of the Shields parameter θ_{50} as:

349
$$\theta_{50(i_e)} = \frac{h \cdot i_e}{\Delta \cdot D_{50}}; \ \theta_{50(i_b)} = \frac{h \cdot i_b}{\Delta \cdot D_{50}} \tag{15}$$

where h is the water depth measured in the field at the monitoring station and D_{50} is 350 the median grain size. Figure 5 shows the typical daily fluctuation of θ_{50} over several 351 days during the monitoring season, calculated with the effective slope i_e . The value of 352 θ_{50} calculated with topographical slope would be one order of magnitude higher than 353 shear stress, which is actually responsible for bed load transport. The outcome using 354 topographical slope suggests that bed load even occurs at the lowest discharges at 355 356 night. In contrast, effective slope provides values of shear stress ranging from 0.030 to 0.075. 357

The shear stress calculated with Eq. 15 can be then applied to the revised MPM equation (Wong & Parker, 2006) to evaluate sediment transport rates. If applied to well sorted mixtures, the MPM reads as:

361

$$q_s^{MPM} = 4(\theta - \theta_{cr})^{3/2} (g\Delta)^{1/2} D_{50}^{3/2}$$
(16)

where q_s^{MPM} represents calculated solid discharge per unit of active channel width. For poorly sorted mixtures, the unit solid discharge can be calculated for each grain size fraction D_i as follows:

365

$$q_{si}^{MPM} = 4f_i(\theta - \xi_i \theta_{cr})^{3/2} (g\Delta)^{1/2} D_i^{3/2}$$
(17)

366 f_i being the volumetric frequency of each i^{th} grain size class and ξ_i the hiding factor 367 calculated as $\xi_i = (D_i/D_{50})^{0.905}$, as previously suggested (Andrews, 1983; Parker et 368 al., 1982). The instantaneous value of q_s^{MPM} is calculated using a value of the Shields 369 stress obtained from Eq. (15), where the flow depth is provided by the local water 370 stage meter.

- 371 Figure 6 shows bed load rates measured using Bunte samplers *vs.* the dimensionless
- 372 shear stress calculated using the effective slope i_e . As expected, bed load rate increases
- with shear stress, rising from 5×10^{-8} m² s⁻¹ at Shields of 0.03 to 5×10^{-5} m² s⁻¹ at Shields
- of 0.06, and also the values of volumetric bed load rates are rather scattered. Figure 7
- shows instead the values of bed load rates *vs*. the water stage at the time of sampling.
- 376 This graph allows plotting the MPM predictions calculated using both i_e and i_b
- 377 showing that using topographic slope to calculate MPM results in overestimating bed
- load transport rates by more than one order of magnitude (Figure 7). Conversely,
- 379 MPM predictive power increases substantially when the effective slope is used,
- 380 especially at the higher transport rates.
- 381 Notably, the application of Rickenmann & Recking's approach (2011) closely
- 382 overlaps with the MPM formula based on i_e . This suggests that the use of effective
- 383 slope is a promising way to increase the predictive power of empirical bed load
- 384 formulas developed in flume experiments, in order to account for energy dissipation
- due to morphological units, such as step-pools, which are usually associated with high
- 386 slope (Rickenmann & Recking, 2011). Indeed, the presence of step-pools produces a
- 387 large apparent increase in Shields stress (Chin & Wohl, 2005; Mao et al., 2008), and
- 388 can strongly affect bed load prediction (Recking et al., 2016).
- 389 It is worth stressing here that the reach-averaged shear stress calculated with Eq. 11 is 390 based on several assumptions (i.e. topographical slope is equal to the water slope and
- to the energy slope) that are hardly met in steep streams like the Estero Morales (see
- 392 Yager et al., 2018). In future attempts, better estimates of effective slope and
- 393 predictions of bed load rates could be achieved by monitoring the slope of the water
- 394 surface or local flow velocity, or by using 2-D flow models to estimate the near-bed
- 395 shear stress (e.g. Monsalve et al, 2016).
- 396 The analysis also provides further insights into the transport regime of Estero Morales
- 397 during summer glacial-melting floods. As previously suggested (Capart & Fraccarollo,
- 398 2011; Berzi & Fraccarollo, 2013), the Shields value can discriminate between ordinary
- 399 (low Shields) and collisional (high Shields) bed load regimes for non-cohesive coarse
- 400 sediments in turbulent flows., Shields stress in the Estero Morales calculated using

401 effective slope is around 0.1, which suggests the occurrence of an ordinary bed load 402 transport regime in which sediment moves by sliding and saltation, with prolonged 403 rest periods, as was observed in the field. In contrast, the application of the 404 topographical slope i_b suggests an intense collisional regime in which particles also 405 frequently collide.

406

407 5. Displacement lengths and virtual velocity of tracers

As previously observed, finer particles in step-pools tend to undergo more 408 409 displacement than coarser particles (e.g. Church and Hassan, 1992). Figure 8 shows the plotted measured total displacement lengths of single tracers scaled by the mean 410 total displacement of the median grain size $(\ell_{50-surf})$ as a function of the ratio 411 412 between tracer size (D) and the median subsurface grain size (D_{50-sub}). The average value of dimensionless flight distance for binned classes of dimensionless particles 413 sizes decreases for coarser sediments. Although considerable scatter is observed, the 414 size-dependent trend is confirmed and the field data are plotted around previously 415 416 derived empirical formulas (i.e. Church & Hassan, 1992; Vasquez-Tarrio et al., 2019). However, the empirical relationship is not statistically significant ($R^2=0.095$; Figure 417 418 8).

Virtual velocity is usually calculated as the ratio between the displacement length and 419 the time between the survey when the tracers were placed in the field and the one 420 421 when they were retrieved. Because the daily fluctuations in discharges in the Estero Morales are fairly regular in summer (see Figure 5), we consider here that daily 422 hydrographs are only characterized by over threshold flows 30% of the time (θ_{50} > 423 θ_{cr}). Following this logic, we calculated the average virtual velocity of each PIT tracer 424 using their travelled distance and 30% of the time interval between the two subsequent 425 426 surveys, irrespective of the survey date.

427 The virtual velocity of particles of different sizes can be used to quantify the

428 volumetric bed load transport rates using Eq. 6, which mainly requires virtual velocity

and the thickness of the active layer δ . The thickness of the active sediment layer

430 generally depends on local hydraulic conditions and bed texture (e.g., Houbrechts et

431 al., 2012). Although δ depends on the shear stress acting locally (Wilson, 1987;

Kumbhakar et al., 2018; Brenna et al., 2019), it can be reasonably assumed that the 432 thickness of the active layer is proportional to the surface grain size $\delta = a D_{50}$ 433 (Haschenburger & Church, 1998) with a ranging from 1 to 3 (Hassan & Bradley, 434 2017). Einstein (1950) originally estimated a as twice the surface grain size. As a first 435 approximation in this paper, we assume that the thickness of the active layer is equal 436 to the median surface grain size. This is reasonable during the ordinary fluctuations of 437 438 discharge monitored in 2014 and 2015, as only a few tracers were found buried during the surveys. Given that the step-pool remained stable over the two years of 439 observations, this indicates that only minor vertical adjustments occurred in the 440 channel bed. Arguably, the assumption that $\delta = D_{50}$ would not hold true for higher-441 442 magnitude events that affect the stability of step-pools sequences, as the thickness of the active layer would become larger and affected by erosion/deposition dynamics. 443 The above considerations on how to evaluate the virtual velocity and the active layer 444 in the field are useful in considering Eq. 7. 445

446 If bed load transport rates q_{si} are calculated using the MPM formula based on the

effective slope (Eq. 17; Figure 7), Eqs. 7 and 17 are combined to obtain V_{vi} in m/day:

$$V_{vi} = \frac{1}{0.3 D_{50}} \int_{1 day} q_{si}^{MPM} dt$$
 (18)

449 Here we assume a value of c_b equal to 1, and a grain size distribution range corresponding to that of the PIT population, thus underrepresenting the finer fractions, 450 to some extent. In Eq. 18 V_{vi} represents the virtual velocity (m/day) averaged over the 451 portion of the day when $\theta_{50} > \theta_{cr}$. The two sides of Eq. 18 are plotted against grain size 452 453 in Figure 9. As documented elsewhere (e.g. Liebault et al., 2012), despite the high degree of scatter, virtual velocities tend to decrease hyperbolically with grain size. 454 This behavior is reproduced fairly well by the MPM volumes calculated when the 455 hiding effect (see Eq. 17) and effective slope are considered. Interestingly, using 456 topographical rather than effective slope in the calculation of bed load transport rates 457 would result in a major overestimation of virtual velocity, especially of the coarser 458 fractions, consistently with results in Figure 7. The overall reduction of virtual 459 460 velocities with grain size indicates that the shear stress exerted by the flow over the 461 bed during a daily flood is never high enough to make the size of transported sediment 462 independent of critical Shields stress (e.g. the transport is size-selective), and confirms

that the sediment transport regime belongs to the ordinary bed load mode.

464 The tracer data show virtual velocities of up to 65 m per day (looking at the binned 465 data in Figure 9) and null values for grain size coarser than 250 mm. These values are 466 representative of the flow conditions obtained by restricting the analysis to intraday 467 sub-periods characterized by $\theta > \theta_{cr}$ (see Figure 5). Clearly, if averaged over an entire 468 day, including periods at $\theta < \theta_{cr}$, when no bed load transport occurs, we would obtain 469 lower virtual velocity.

470

471 **6.** The statistics of sediment resting times and Einstein conjecture

Ordinary tracer studies focus on the mobility and virtual velocity of sediments during 472 473 floods. Frequent subsequent surveys with the mobile antenna, combined with the natural daily fluctuations in discharge, allowed us to focus on the tracers that remained 474 in the same location over two or more surveys in order to calculate sediment resting 475 times. The field surveys conducted over consecutive days allowed for directly 476 measuring the resting times T_i of the PIT tags of different grain sizes. When a tracer 477 was recovered in the same position in the following survey, the resting time was 478 calculated, representing the time interval elapsing from the first to last retrieval in the 479 same position. Using these direct measurements, the mean resting time obtained by 480 481 averaging all the PITs is approximately 8.6 days. If the resting times are calculated using the portion of the day when $\theta > \theta_{cr}$ (which represents approximately 30% of the 482 time, see Figure 5), the mean resting time drops to around 2.6 days of continuous over-483 484 threshold flow with active bed load transport. Figure 10 reports the values of these directly measured resting times (red dots) as a function of tracer size. The lack of data 485 points for durations of less than 8 hours (30% of a day, when $\theta > \theta_{cr}$) is related to the 486 temporal resolution of field surveys (i.e., at least one day apart from each other), 487 488 which is an intrinsic limitation of non-continuous tracking. Interestingly, the resting times are not sensitive ($R^2 = 0.03$) to the grain size of tracers, indicating that the 489 probability of a grain in the Estero Morales being entrained is independent of its size 490 (Figure 10). 491

Figure 10 also shows a series of resting times calculated indirectly, derived from a re-evaluation of the Einstein approach along with the empirical resting times derived

494 from repeated field surveys. The method used to calculate this dataset is explained495 below.

496 We took advantage of the fact that any measured PIT displacement is the cumulative distance travelled over a known inter-survey time interval, without taking into account 497 that particles tend to move on subsequent jumps that include multiple stops with 498 relative resting times. Conceptually, at least one resting time is associated with any 499 500 single flight (i.e., a distance > 0) that took place during an inter-survey interval. We developed a method to work with these data and extract one or more resting time from 501 any inter-survey displacement. In Einstein's (1950) bed load formula (Eq. 2), the flight 502 distance \bar{L}_i is assumed to depend on the size of the sediment D_i and on ω , a positive 503 dimensionless constant ($\omega \approx 100$): 504

505

$$\bar{L}_i = \omega D_i \tag{19}$$

Given that \bar{L}_i depends only on D_i (see Eq. 19), in Einstein formula (Eq. 1) the bed load intensity depends completely on the other predictor E_i representing the number of particles entrained per unit time and unit bed-surface area. In order to interpret our dataset, we relaxed the Einstein constraint and adopted the following linear relationship:

511

$$\bar{L}_i = \omega D_i + c \tag{20}$$

The constant *c* makes it possible, at least theoretically, to obtain a negative value of the coefficient ω , which is always positive in experimental studies (Campagnol et al., 2012). The presence of *c* also implies that there are parts of the bed surface occupied by stable bed-form structures over which particles do not stop during ordinary events (i.e., floods are not able to destroy bed forms). As in Einstein (1950), Eq. 20 assumes that the coefficients are constant and independent of both grain size and bed load intensity.

519 By applying a recursive procedure, we obtain L_i and a rich sample of resting times that 520 together contribute to describing the distribution of the directly measured resting 521 times. In the first iteration of the procedure, which started with guess values of ω and c522 (e.g., the Einstein's values, $\omega = 100$, c = 0 m), and we used Eq. 20 to calculate L_i for 523 all grain sizes. Each measured inter-survey displacement was divided by L_i to obtain 524 the number of resting periods the particle underwent during such displacement. The

inter-survey time interval was thus divided by the number of resting periods to 525 calculate the duration of each one. These additional and indirect values enrich the 526 527 statistics of the resting times, summing to those directly measured in the field. All the values of resting times, either directly measured and inferred from the 528 displacements, were used to calculate a mean value \overline{T} . We then used the measured 529 average values of virtual velocity for each grain size (see Figure 9) to indirectly obtain 530 the average flight distance L_i based on the grain size. This is accomplished by 531 multiplying the virtual velocity, which depends on the grain size (we used the binned 532 values reported in Figure 9), by \overline{T} . Based on these values of L_i the parameters ω and c 533 were back-calculated, using Eq. 20, and iterations were repeated until the values of ω 534 and c remained constant. 535

536 Figures 10 and 11 illustrate the results obtained at the end of the direct and indirect approaches of the calculation. Neither directly measured data (red dots in Figure 10) 537 538 nor the indirect data extracted from displacements (blue dots in Figure 10) show significant dependence of resting times on grain size. Figure 11 shows the statistical 539 540 distribution of resting times of the two datasets. The directly measured resting times displays a shapeless probability-density distribution (Figure 11) which becomes much 541 more outlined when measured and displacement-inferred resting times are merged 542 (Figure 11b). Their overall mean is about 32 hours (1.35 days). The directly measured 543 and the displacement-inferred resting times concur when below-threshold periods (i.e. 544 when the Shields value is below the critical level) have been removed. Because θ 545 exceeds θ_{cr} for approximately 30% of the time, an effective mean period T_r of 32 hours 546 represents on average 4.5 days in the monitored reach of Estero Morales. Figure 11 547 clearly indicates that the inequality reported in Eq. 3 fully applies, as the probability 548 density distribution of the resting times are far from being uniform. 549 Our results further show that the relation between L_i and D_i does not confirm Einstein's 550 assumption that $\omega = 100$ and c = 0 m. Indeed, the flight distance of single grains 551 decreases with particle size, and we obtained values of ω and c of -200 [-] and 50 m, 552 respectively (Figure 12). Interestingly, ω is negative, revealing that in a naturally 553 structured step-pool channel experiment, sediment mobility is completely different 554 from that in flume runs with well sorted sediments and no bedforms (Campagnol et al., 555

2012; Lajeunesse et al., 2010; Lee et al., 2006; Radice et al., 2013; Fraccarollo &
Hassan, 2019). Deviations from Einstein conjecture were reported by Habersack
(2001) in one of the first reports on step lengths and rest periods using tracers in the
field.

560 It is worth discussing the reason ω has a negative value. Once small particles are dislodged, they are transported for long distances, before coming to rest. This is 561 562 consistent with what is shown in Figure 10 and with the outcomes of recent works (e.g., Vázquez-Tarrío et al., 2019). The value obtained for c is higher than the average 563 564 distance between steps in Estero Morales, which is around 10 m. The spacing between pools is a relevant geomorphic dimension of step-pool sequences, given that it is 565 566 closely related to step formation (Curran, 2007) and flow resistance (Giménez-Curto & Corniero, 2006). During a given ordinary glacier-melting induced flood, the tracer 567 568 displacements are influenced by the bed morphology and selected according to the grain size. Pools are particularly difficult to pass, and it is where coarse particles are 569 570 more likely found. The work by Habersack (2001) is one of the few to have dealt with Einstein predictors using field data. This author radio-tracked gravel particles in a 571 large braided river and gathered hundreds of direct measurements of resting times and 572 flight distances. Interestingly, although working in a different bed form setting (i.e., 573 dunes, gravel sheets and bars), Habersack (2001) also found a negative relationship 574 between the dimensionless step lengths L_i/D_{50} and scaled particle size (D_i/D_{50}) . 575 Through flume experiments, Pyrce and Ashmore (2003) showed that the distribution 576 of mean flight length is related to pool-bar spacing and that only a small proportion of 577 particles are able to move beyond the first bar downstream from the entrainment site. 578 579 Present and previous results confirm the dependence of grain flights on morphological 580 features, in particular the control that bed forms can exert on coarse particles, preventing long flights, which support the need for a negative value for the 581 proportionality parameter ω in our generalization of the Einstein relation (Eq. 20). 582 583 It is worth stressing that all results presented here derive from field measurement 584 conducted during ordinary daily floods generated by the glacial melt. Larger events 585 (i.e. intense rainfall events in early autumns in the case of the Estero Morales) would 586 result in disruption of the step-pool morphology, deeper sediment active layer, and

587 large erosion/deposition processes in the channel and banks. Under these conditions,

the statistical features of sediment displacement lengths and resting times would be

589 likely quite different, and the very premise of the present analysis, i.e. the

590 decomposition of the topographic slope, would be affected. It would be interesting to

- 591 corroborate the Einstein's contention under higher magnitude events in the Estero
- 592 Morales when, as in laboratory flume experiments, most of the bed would be mobile.
- 593

594 **7. Conclusions**

In this work, we presented and analyzed a new dataset collected over two years of 595 sediment transport monitoring, including PIT-tracer displacements and direct 596 597 volumetric sampling with Bunte traps, in Estero Morales (Chile). The latter data allowed us to assess the effective slope of the study reach, at least for ordinary floods, 598 and to apply the relation proposed by Meyer-Peter and Müller to successfully predict 599 sediment discharge. This physically based method of breaking down the total (or 600 601 geometric) slope yielded results that are in good agreement with previous formulations and can be tested straightforwardly in other streams, provided a meter gauge is 602 603 installed in the reach and volumetric samples of bed load are taken in a range of discharges. We further presented the link between Einstein's predictors and virtual 604 605 velocity. Frequent tracking of tracer positions, along with the regularity of daily floods 606 in Estero Morales, allowed us to collect a first sample of resting times. Then, by adopting a generalized version of Einstein's fundamental law, we inferred additional 607 608 resting times and flight distances from inter-survey data of tracer displacements. Our results contradict Einstein's assumption and show a surprising decrease in flight 609 610 distances with larger grain sizes, as opposed to a linear increase. This finding is 611 conceptually consistent with the notion that bed macroroughness, which is responsible 612 for slope decomposition, plays a prominent role in processes that characterize sediment transport, like entrainment and deposition. We also observed that resting 613 614 times do not depend significantly on tracer size.

615

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- online open data repository (DOI: 10.5281/zenodo.3274762).
- 625

626 FIGURE CAPTION

- 627 **Figure 1.** Location and map of the Estero Morales basin.
- **Figure 2.** Views of the Estero Morales (a) at the monitored cross-section; (b) at the
- 629 upstream end of the reach where the PIT tracers were placed and surveyed; (c) during
- 630 a PIT survey conducted with a portable antenna; and (d) during bed load sampling
- 631 using Bunte traps. The channel width is about 7 m.
- 632 **Figure 3.** Surface and sub-surface grain size distribution in the monitored reach of the
- 633 Estero Morales, and size distribution of PIT tracers.
- **Figure 4.** Relationship between the coarsest clast (D_{100}) collected with the Bunte
- 635 sampler and the water depth at the time of sampling, as measured at the monitoring
- 636 site during events recorded in 2014 and 2015). The filled black dots represent the
- 637 average values of D_{100} for binned classes of water depth. The regression plots as D_{100}

 $638 = 0.2436h - 0.0056 \ (R^2 = 0.29).$

- **Figure 5.** Temporal trend of the dimensionless shear stress over three exemplary
- 640 weeks of January 2014 in the Estero Morales calculated with Eq. 15 using the
- 641 effective slope i_e . The horizontal line refers to the assumed critical value of shear
- 642 stress ($\theta_{50(ie)} = 0.05$), which corresponds to a value of Shields of almost 0.40 if
- 643 calculated using the topographical slope i_b .
- **Figure 6.** Bed load transport rates as measured with the Bunte traps in 2014 and 2015,
- 645 plotted against the Shields parameter calculated using the effective slope. The filled
- black dots represent the average values of q_s for binned classes of shear stress.
- 647 **Figure 7.** Bed load transport rates plotted against the water depth water depth at the
- 648 time of sampling. The filled black dots represent the average values of q_s for binned

- 649 classes of water depths. The figure shows that the Meyer-Peter Mueller equation
- 650 predicts the order of magnitude of bed load rates if the Shield stress is calculated using
- 6_{51} i_e , but overestimates actual transport rates by more than two orders of magnitude if
- 652 calculated using i_b .
- 653 Figure 8. Dimensionless displacement ℓ_* of individual particles (scaled by the
- 654 averaged travel distance of particles similar to the surface size D_{50}) as a function of
- dimensionless particle size D^* (scaled by the subsurface median size). The best fit
- 656 regression is $\ell_* = -0.296 D_* + 2.389$ (n=517; P<0.001; R²=0.095). The filled dots
- 657 represent the average values of dimensionless displacements for binned classes of
- dimensionless particles sizes. Empirical equations of Church & Hassan (1992) and
- 659 Vasquez-Tarrio et al. (2019) are plotted as well.
- 660 Figure 9. Virtual velocities plotted against the size of tracers used in the Estero
- 661 Morales. The filled dots represent the average values of virtual velocity for binned
- classes of particles sizes. The solid line represents Eq. 17.
- **Figure 10.** Resting times of coarse particles directly measured or computed through
- 664 displacement surveys of PIT tags.
- **Figure 11.** Histograms of the rest periods from direct surveys (left panel), and all rest
- 666 periods including those calculated through displacement surveys (right panel).
- 667 Figure 12. Experimental data of flight displacement versus grain size, including the
- best fit of Eq. 20. Einstein's line represents $L_o = 100D_i$. The empirical equation of
- Habersack (2001), applied using D_{50} of the Estero Morales is plotted as well.
- 670

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Figure 1.



Figure 2.



Figure 3.



Figure 4.



Figure 5.



Figure 6.



Figure 7.



Figure 8.



Figure 9.



Figure 10.



Figure 11.



Figure 12.

