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Key Points:

- The disequilibrium ratio represents the observed bedload mass per event divided by the integrated bedload transport capacity
- Disequilibrium transport ratio and threshold transport discharge are inversely related to each other and vary in cycles
- Increased sediment supply or availability results in longer cycles, a larger active layer depth, and smaller threshold transport discharge

Supporting Information:

- Supporting Information S1

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Effect of Sediment Supply on Cyclic Fluctuations of the Disequilibrium Ratio and Threshold Transport Discharge, Inferred From Bedload Transport Measurements Over 27 Years at the Swiss Erlenbach Stream

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Abstract Bedload transport in natural channels typically shows large fluctuations even for constant hydraulic forcing. It has been shown that the threshold shear stress for initiation of motion depends, for example, on the bed morphology and on channel slope and that it may be influenced by flood history effects. This study is based on a unique data set of continuous field observations of bedload transport acquired during 27 years with impact plate measurements and involving more than 500 sediment-transporting flood events. A disequilibrium ratio is defined which represents the observed bedload mass per event divided by the bedload mass calculated via a bedload transport equation integrated over the time of a flood event. The disequilibrium ratio varies in a cyclic behavior around equilibrium conditions defined by its median value. The mean duration of a cycle depends on upstream sediment supply or availability on the bed, and longer cycles are associated with a larger active layer on the bed. A memory effect is evident regarding the temporal evolution of both disequilibrium ratio and bed state as reflected by threshold transport conditions, determined from the observations as the critical discharge at the start and at the end of a bedload-transporting event. The disequilibrium ratio and the threshold transport discharge are inversely correlated with each other, likely providing a feedback mechanism governing the fluctuations around an equilibrium state. Accounting for the memory effect, that is, for the bed state after the previous event, improves the prediction of bedload transport for the following event.

1. Introduction

Measured transport rates in gravel-bed rivers and boulder-bed streams vary by several orders of magnitude for given (mean) flow conditions (e.g., Bathurst et al., 1987; Gomez & Church, 1989; Rickenmann, 2016; Whitaker & Potts, 2007a). Recking et al. (2016) demonstrated, based on more than 10,000 field measurements, that the reference shear stress varies as a function of channel slope and that the steepness of the bedload transport function showed some dependence on the dominant bed morphology, such as plane bed, riffles, step pools, or braided morphology. In naturally occurring flows, the variability of grains and their local arrangement affects the hydrodynamic and resisting forces, and thus, the net hydrodynamic force in turbulent flow, represented by the sum of the lift and drag contributions, will vary in time, magnitude, and direction (Diplas et al., 2008). A coarsening of the bedload with increasing flow strength was observed in several gravel-bed streams (e.g., Bathurst, 1987; Ryan et al., 2005; Whitaker & Potts, 2007b). It appears therefore that one reason for the bedload transport variability is related to changes in bed evolution affecting the critical Shields stress (dimensionless shear stress for the start of bedload transport), such as changes in packing, grain size of the bed surface, roughness, imbrication and orientation of the particles, clusters, and armor layer configuration (Church, 2006; Diplas et al., 2008; Guney et al., 2013; Haynes & Pender, 2007; Humphries et al., 2012; Mao, 2012; Piedra et al., 2012; Roth et al., 2014, 2017; Waters & Curran, 2015).

A second reason for the transport variability is the stabilization of the bed surface during low-flow periods without bedload transport (interevent flows) and during flow periods with only weak transport, leading to an increase in critical Shields stress or a reduction in sediment transport over time (Haynes & Pender, 2007; Mao, 2018; Masteller & Finnegan, 2017; Monteith & Pender, 2005; Ockelford et al., 2019;

Ockelford & Haynes, 2013; Paphitis & Collins, 2005; Reid et al., 1985). Recently, Masteller et al. (2019) analyzed 19 years of continuous bedload transport observations at the Swiss Erlenbach stream, and they observed a strong dependence of the critical Shields stress on the magnitude of previous flows. They found that channel stabilization occurred during times with weak and medium bedload transport and times without transport, and they suggested that local rearrangement of grains is responsible for this memory effect on the critical Shields stress.

A third reason for the transport variability is sediment supply and sediment availability on the bed surface. Exceptionally high discharge events can reduce particle interlocking through boulder mobilization and the breaking of channel bed structures (Turowski et al., 2009; Yager et al., 2012), resulting in increased bedload transport rates (Lamarre et al., 2008; Oldmeadow & Church, 2006; Yager et al., 2012). Higher rates of upstream sediment supply tend to result in more mobile sediment on the bed and thus in lower critical Shields stresses (Bunte et al., 2013; Dietrich et al., 1989; Recking, 2012). Masteller and Finnegan (2017) found an increase in the number of highly mobile, high protruding grains in response to sediment transporting flows in flume experiments, implying an associated decrease in the critical Shields stress. Overall, the cited observations suggest that gravel beds integrate the effects of both interevent flows and bedload transport events by reorganizing the grains and the structure of the bed surface, resulting in a history-dependent Shields stress (Masteller et al., 2019).

A state-dependent model for bedload transport was proposed by Johnson (2016), in which the critical Shields stress depends on the history of bedload transport and evolves as a power law function of net erosion or deposition. The model theory is inspired by thermodynamics, and channel morphology (Shield stress) is considered to be a bed state variable (similar to temperature). The cumulative discharges of both water and sediment are considered to be path variables (similar to heat) that drive bed state evolution. During periods of net entrainment, mobile particles are expected to be removed progressively, gradually increasing critical Shields stress. During periods of net deposition, topographic lows along the bed are filled with particles, progressively leading to less stable distributions of surface grains and thus decreasing critical Shields stress. Johnson (2016) tested his model with flume experiments to explore systematic feedback effects on disequilibrium gravel transport. Chartrand et al. (2019) used scaling theory to develop two dimensionless channel response numbers which are a measure for the local disequilibrium state of gravel-bed mountain streams. The dimensionless numbers express a balance between (i) the rate of local bed topography adjustment and (ii) the rate of bed sediment texture adjustment.

Long-term measurements of bedload transport in natural channels may help to better test existing models concerning the effect of bed state (morphological structures, grains size distribution, interlocking of particles, and roughness), discharge (flow strength), and sediment supply on bedload transport fluctuations under field conditions. Surrogate bedload transport measuring techniques with acoustic or seismic sensors offer a possibility for such observations, particularly in steep gravel-bed streams (e.g., Gimbert et al., 2019; Rickenmann, 2017). However, so far, only a few long-term data series of fluvial bedload transport are available (Aigner et al., 2017; Downs et al., 2016; Habersack et al., 2017; Kreisler et al., 2017; Rickenmann, 2018; Rickenmann & McArdell, 2007, 2008; Schneider et al., 2016).

In this study I analyzed the temporal evolution of bedload-transport fluctuations over more than 500 flood events which occurred in the Swiss Erlenbach stream between 1986 and 2016, where bedload transport has been monitored continuously with impact sensors. Along with continuous discharge measurements, the monitoring data allowed to determine the threshold conditions at the start and at the end of a sediment-transporting flood event, as well as the bedload transport capacity. Observed event bedload mass and event-integrated bedload transport capacity allowed to define equilibrium transport conditions for the given range of event magnitudes. One major objective of this study was to search for similar patterns characterizing fluctuations around a bedload transport equilibrium as proposed by Johnson (2016). More specifically, I used correlation analysis (i) to explore the degree of interdependence between key variables such as measured event-bedload mass, event-integrated bedload transport capacity, and threshold discharge and (ii) to identify systematic trends in the fluctuations of key variables around an equilibrium or mean condition and any existing memory effects over the course of the events. Furthermore, I combined the observations at the bedload-transporting monitoring cross section with findings from previous Erlenbach studies to (iii) examine how the observed fluctuations around equilibrium conditions are related to upstream channel reach sediment supply and active layer thickness.

2. Field Site and Data Analysis

2.1. The Erlenbach Catchment

The Erlenbach catchment is located in the Prealps of central Switzerland (Hegg et al., 2006). The catchment has an area of about 0.74 km², and the stream gradient is 18% on average. Continuous measurements of rainfall, runoff, and sediment transport have been taken since 1986 (Hegg et al., 2006). The hydrology of the catchment is characterized by both frequent high-intensity storms in the summer and snowmelt events in the spring. The Erlenbach channel has a step-pool morphology with some riffle and cascading reaches and an average bankfull channel width of 3.7 m (Molnar et al., 2010). Bedrock outcrops are very rare along the channel. Most of the catchment is located on a large landslide complex, and the left bank is particularly active in supplying sediment to the channel by hillslope creep in the reach upstream of the gauging station (Golly et al., 2017; Schuerch et al., 2006). Hillslope landslides are very active, and they have the potential to supply considerable amounts of sediment to the channel and out of the basin (Molnar et al., 2010). A study reach of about 550 m length upstream of the sediment retention basin was repeatedly surveyed for the longitudinal profile. This reach has a mean gradient of about 15% and includes about 80 steps, with a mean step height of about 0.7 m (Molnar et al., 2010).

A minimum discharge of about 7 to 10 m³/s must be exceeded to destroy many or most of the steps (Turowski et al., 2013). Since the start of the continuous bedload transport measurements in 1986, only three flood events occurred (1995, 2007, and 2010) that had a peak discharge of about 10 m³/s or more. After such events, maximum vertical erosion into the streambed reached locally between 2 and 3 m (Molnar et al., 2010; Turowski et al., 2009, 2013). The steps can reform relatively quickly after exceptional flood events. The term “exceptional” is used here sensu Lenzi et al. (1999), that is, floods with a large and lasting effect on stream morphology (see also Turowski et al., 2009).

2.2. Bedload Transport Measurements

Near the catchment outlet, there are two stream gauges, a sediment retention basin, and steel plates with impact sensors to continuously measure bedload transport since 1986. Between 1986 and 1999, piezoelectric bedload impact (PBI) sensors were used. These sensors record the vibrations resulting from the movement of gravel-sized and larger particles over a steel plate (Rickenmann & McArdell, 2007). After deterioration of the homemade PBI sensors had begun, they were replaced by geophone sensors from 2000 onward (Rickenmann, 2017). The bedload measurements with both sensors give a very similar signal response (Rickenmann et al., 2012). The minimum grain size that can be detected by both systems is about 10 mm (Rickenmann et al., 2012, 2014; Rickenmann & McArdell, 2007; Wyss et al., 2016a). The steel plates equipped with an impact sensor are mounted flush with the streambed, just upstream of the sediment retention basin (Figure 1a). The steel plates have standard dimensions of $L \times B \times T = 358 \times 496 \times 15$ mm, where L is the downstream length, B is the transversal width, and T is the thickness of the plate.

Regarding the calibration of the surrogate measuring system, I use here the summary value of the impulse counts: Whenever the voltage of the raw signal exceeds a preselected threshold value A_t (in V), this is recorded as an impulse, and the summed impulses are stored. The use of a threshold essentially eliminates the noise of the signal. A linear relation between impulses and bedload mass transported over a plate was found to apply quite well at several field sites, including the Erlenbach (Rickenmann et al., 2012, 2014; Wyss et al., 2016a, 2016b). For the Period A with the PBI sensors (1986–1999), a total of nine plates was equipped with a sensor. The lateral distribution of these sensors is illustrated in Figure 1b. An asymmetrical lateral distribution of bedload transport intensities was observed, which is due to an offset of the centerline axes of the approach flow channel with regard to the position of the check dam upstream of the retention basin. Therefore, most signal was recorded by Sensor H3, followed by Sensors H4 and H9 (Rickenmann & McArdell, 2007). For the Period B with the geophone sensors (2002–2016), a total of six plates was equipped with a sensor. The lateral distribution of the geophone sensors is illustrated in Figure 1b. A similar asymmetrical lateral distribution of bedload transport intensities was found also for this period. The threshold value to record impulses was set to $A_t = 0.2$ V for the PBI and to $A_t = 0.1$ V for the geophone sensors. The summed impulse counts IMP were recorded in minute intervals during bedload-transporting flood events.

The linear calibration relation used to convert the number of impulses into a channel-wide bedload transport rate Q_b , including grains larger than 2 mm for both periods, is the following:

Figure 1. (a) Locations of the surrogate bedload impact plates, the sediment retention basin, and the two gauging sites at the Erlenbach (near the catchment outlet). (b) Downstream view of the lateral distribution of PBI sensors (H_i) during Period A (1986–1999) and of geophone sensors (GPi) during Period B (2002–2016), where i denotes the sensor number. (c) Upstream view of the symmetrical, trapezoidal cross section used for the hydraulic and bedload-transport calculations, with a bottom width $w = 4.1$ m and a lateral slope of 1.5:1; this cross section represents the natural reach upstream of the measuring installations where the banks are protected by a riprap construction (inset photo).

$$Q_b \text{ (kg/s)} = k_{isp} \text{ (IMP/1000)} f_g, \quad (1)$$

where k_{isp} are the individual calibration coefficients determined for each survey interval of the deposits in the sediment retention basin. The lumped factor $f_g = 22.46$ in Equation 1 accounts for the mean bulk density of the deposits in the sediment retention basin, the estimation of the proportion of particles larger than 2 mm in the deposits, and for the conversion of the 1 min IMP readings to a mean value of Q_b expressed in (kg/s). More details on the calibration relation are given in the supporting information both with regard to calculation of the factor f_g (Text S1) and the calibration coefficients k_{isp} (Table S1).

2.3. Discharge Measurements and Sediment-Transporting Flood Events

There are two runoff gauging stations near the sediment retention basin, which is close to the outlet of the catchment. The so-called lower station at the outflow from the (dammed) sediment retention basin consists of a double triangular profile; the associated stage-discharge relationship (runoff Q_u) was calibrated with a physical model before construction and is therefore quite reliable also during flood flows. The upper discharge measurement station is located between the end of the natural channel and the retention basin, and it consists of an asymmetrical cross section in a concrete channel; the associated stage-discharge relationship (runoff Q_o) was calibrated based on dye and salt tracers measurements, especially for small and medium discharges but including discharges up to $5 \text{ m}^3/\text{s}$. The location of the two gauging sites is illustrated in Figure 1a and in Beer et al. (2015, Figure 2).

For the past sediment-transporting flood events, the discharges were in the range of 0.1 to $15 \text{ m}^3/\text{s}$ (Rickenmann, 1997; Turowski et al., 2009). During flood events, level measurements are available in a time interval of 1 min for both stream gauging sites. The definition of a sediment-transporting flood event is based on the recording of the bedload transport activities. More details on the bedload transport criteria to start and end the recording of an event are given in Text S2.

On some days, bedload transport activity ceased for minutes to hours and restarted again. A typical duration of a sediment-transporting flood event varies between about 20 and 300 min, and a mean duration for the thunderstorm-triggered events during summer is about 100 min (Rickenmann, 1994). In some cases, for example, if rainstorm intensity decreased and increased again within minutes or hours, the runoff remained slightly below the threshold discharge for a limited time and generated a hydrograph with two peaks separated by a few hours. Therefore, I defined a minimum interflood duration time t_c (without any transport

Table 1
Study Periods, Bedload Sensors, and Number of Sediment-Transporting Flood Events

Period	Years	Bedload sensor	No. of flood events for $t_c = 75$ min	No. of flood events retained, with $M_{\text{gravel}} > 50$ kg and $MQ_{\text{bt}} > 20$ kg	Mean no. of flood events per year
A	Oct 1986 to Sept 1999	PBI	318	286	22.0
B	Nov 2002 to Oct 2016	Geophone	263	236	16.9

signal) to separate different flood events. I varied t_c values between 30 and 90 min and compared the resulting number of events with a visual or “manual” identification and separation of flood events. The best agreement was found for a value of $t_c = 75$ min, which is of the order of the duration of a typical flood event. For both Observation Periods A and B together, this resulted in 581 flood events for a total period of 27 years (Table 1).

Regarding the uncertainty of the level-discharge relationship, for some (limited) periods, the Q_o values from the upper gauging station had to be replaced by Q_u values from the upper gauging station. We estimated a similar uncertainty of 15% for the Q_o values (as estimated by Beer et al., 2015) or somewhat less. More details on the discharge measurements are given in Text S3.

Immediately downstream of the upper gauging site, the flow drops over an engineered overfall structure (~ 1 m high) into a shallow stilling basin (~ 4 m $\times$ 4 m $\times$ 0.3 m depth) and then enters a fairly smooth ~ 30 m engineered concrete reach with embedded riprap before reaching the impact plates (Roth et al., 2017, Figure 1 therein). This stilling basin somewhat delays the sediment transfer between the gauging site and the impact plates. Assuming that a temporary storage of bedload occurs in a part of the stilling basin (~ 2 m $\times$ 1 m $\times$ 0.1 m depth = 0.3 m³), a porosity of 0.6, a sediment density of 2,650 kg/m³, and a substantial delay of 10% of the temporarily stored bedload at smaller discharges, this results in a potentially delayed delivery of about 50 kg of bedload mass. This mass corresponds to about 40 impulses for Period A and about 53 impulses for Period B. Note also that the criterion to start and stop an event recording (Text S2) corresponds to a bedload mass of about 25 kg for Period A and about 19 kg for Period B. These considerations provide an estimate of the uncertainty associated with the delineation of single sediment-transporting flood events based on the bedload transport monitoring.

Integrating the observed bedload transport rates Q_b , based on the Calibration Equation 1, over the duration of a flood event, gives the observed bedload mass M_{gravel} for each flood event. Using hydraulic (mostly based on Q_o) and bedload-transport calculations, I have determined the (expected) event bedload masses MQ_{bt} for each flood event (section 2.4 and Appendix A). This resulted generally in a fairly good correlation between the two variables. Only very small values of M_{gravel} and MQ_{bt} did not follow the general trend defined by the rest of the data. Assuming that the Q_o values underestimate the true discharge by 10% at values of $Q_o = 0.5$, 0.6, and 0.7 m³/s, this results in an underestimation of MQ_{bt} by 4, 14, and 32 kg for a 1 min interval. Given the uncertainties of the bedload transport and the discharge measurements, I decided to retain only those events for the further analysis which had a value of both $M_{\text{gravel}} > 50$ kg and $MQ_{\text{bt}} > 20$ kg. This reduced the total number of flood events by 10% for both Periods A and B (Table 1), but it diminished the total bedload mass only by 0.12% for Period A and by 0.21% for Period B.

Based on the discharge measurements and the delineation of sediment-transporting flood events, three further variables were determined which characterize these flood events. $Q_{b,s}$ (in m³/s) is the threshold discharge at the start of an event, $Q_{b,e}$ (in m³/s) is the threshold discharge at the end of an event, and Q_{max} (in m³/s) is the peak discharge observed during an event; most of these values are based on the monitoring at the upper gauging station (Q_o), except for some cases with missing or implausible values (in this case they were based on the Q_u values). According to definition of the beginning of an event described above, the definition of the threshold discharge at the start and the end of an event is objective in the sense that there is a clearly defined criterion (nonzero number of impulses), which is independent of an observer. This criterion corresponds to a small transport rate, and it is thus comparable to other methods to define the critical Shields number as discussed in Buffington and Montgomery (1997).

2.4. Hydraulic and Bedload-Transport Calculations

The hydraulic and the bedload-transport calculations were made for the measured trapezoidal cross section in the natural reach upstream of the measuring installations close to the retention basin (Figure 1c). From four longitudinal profile surveys conducted in 1993, 1995, 2004, and 2007, a mean channel slope was determined for a 20 m long reach upstream of the trapezoidal cross section. This resulted in a mean channel bed slope of $S = 10.5\%$ (varying between 10.0% and 10.8%), which was selected for the calculations. Different values were reported in previous studies for this lowermost study reach of Erlenbach, varying from 9.8% up to 12% (Nitsche et al., 2012; Schneider et al., 2015; Yager et al., 2012).

Characteristic grain sizes in this reach may also have varied over time; however, no regularly repeated surveys are available. The hydraulic calculations are based on dye tracer measurements reported by Nitsche et al. (2012), and I used the value of $D_{84} = 0.29$ m for the Erlenbach bed surface material reported therein. For the bedload-transport calculations, I used the value of $D_{50} = 0.06$ m for the Erlenbach bed surface material reported in Schneider et al. (2015). D_{xx} is the grain size for which xx% of the particles are finer.

The hydraulic calculations were made with a hydraulic geometry flow resistance relation developed by Rickenmann and Recking (2011), which was verified with field measurements in the Erlenbach by Nitsche et al. (2012). The bedload-transport calculations were performed with an equation reported in Schneider et al. (2015), which represents a modified form of the Wilcock and Crowe (2003) equation and which predicts total bedload transport rates (not fractional transport rates) for grain sizes larger than 4 mm. This bedload-transport calculation is based on total shear stress, and integrating the transport rate for the entire channel width over the duration of the flood event resulted in the calculated or “predicted” bedload mass MQ_{bt} (in kg) for each flood event. Further details of both hydraulic and the bedload-transport calculations are given in Appendix A.

2.5. Definition and Calculation of Transport Disequilibrium Ratio

The use of a bedload transport equation (as described in section 2.4 and in Appendix A) resulted in the calculated bedload mass MQ_{bt} for each flood event which was compared with the observed bedload mass M_{gravel} (as described in section 2.2), with a trend for underestimating the M_{gravel} values for smaller events (Figure A1). Using a power law regression between M_{gravel} and MQ_{bt} to correct for this bias resulted in an improved estimate of the event bedload mass, M_{greg} , which is still based on the integrated transport capacity (shear stress) over the flood event:

$$M_{greg} = 18.65 MQ_{bt}^{0.69}. \quad (2)$$

This improved estimate of the event bedload mass shows a similar deviation between observed and calculated bedload masses for the entire range of event magnitudes. Equation 2 defines an (empirical) equilibrium condition in terms of transported bedload mass and hydraulic forcing for the entire range of observed event bedload masses (including 27 years of measurements). The ratio of observed to calculated bedload mass is therefore considered as a disequilibrium ratio, E_d :

$$E_d = M_{gravel}/M_{greg}. \quad (3)$$

The ratio E_d can be thought of alternatively as a measure of the solid-liquid ratio of the flow. The higher this value is, the larger is the solid concentration for a given (event-averaged) hydraulic forcing.

The definition of this disequilibrium ratio resembles in some way the definition of Bagnold's (1966) bedload transport efficiency $e = i_b \tan(\phi)/\omega$, where i_b = immersed unit bedload transport rate, ϕ = dynamic friction angle of grains under water, and ω = unit stream power. The unit stream power ω is proportional to the boundary shear stress τ in the form of $\omega \propto \tau^{1.5}$ (e.g., Shih & Diplas, 2018), and there is the same proportionality in the relation between unit bedload transport rate q_b and τ in the form of $q_b \propto \tau^{1.5}$ for the high-intensity transport region in many bedload transport equations (as, e.g., also in Equation A4b). Concerning the hydraulic forcing, it may be noted that, in a study of developing a rational sediment transport scaling relation, Eaton and Church (2011) concluded that using either dimensionless unit stream power or dimensionless shear stress yields very similar results, except that the latter additionally requires the additional use of a

Figure 2. Comparison of observed bedload mass M_{gravel} for each flood event and calculated bedload mass M_{greg} , using Equation 2. The black line is a power law regression between the two variables, and the red dotted line is the 1:1 line.

Figure 3. Pearson correlation coefficients R calculated for variables characterizing the sediment-transporting flood events. The R values were calculated in the log domain for all variables. All correlations of cells with a nonwhite background are significant with p values larger than 0.05. (The calculations are based on 522 events as indicated in Table 1.)

flow resistance equation. Thus, the transport efficiency e and the disequilibrium ratio E_d will vary in a similar way for flows in the high-intensity transport region or for events where bedload transport is dominated by flows in this transport region. However, back-calculated transport efficiencies e generally decrease for decreasing shear stresses smaller than about the reference shear stress (Abrahams & Gao, 2006; Eaton & Church, 2011; Recking, 2012; Shih & Diplas, 2018), and thus, e and E_d will clearly differ in these flows. It is noted that in some other publications, the term sediment transport efficiency was used not in the strict sense of Bagnold's definition but in a more loose, relative sense, mostly in the way of comparing sediment (bedload) transport with hydraulic forcing for constant flow conditions or integrated over a flood event time scale (Piton & Recking, 2017; Rainato et al., 2017; Recking et al., 2012; Richardson et al., 2020). I have used here the disequilibrium ratio E_d in a very similar way.

3. Results

3.1. Correlations Between Key Variables Characterizing Hydraulics and Bedload Transport

The improved estimate of the event bedload mass, M_{greg} , from Equation 2 shows a fairly good correlation with the observed event bedload mass, M_{gravel} (Figure 2). According to the correlation matrix in Figure 3, there are also significant correlations between the threshold discharge $Q_{b,e}$ and some other variables describing bedload transport (M_{greg} , E_d) and flow strength (MQ_{bt} , Q_{max}). A fairly good correlation can also be observed between $Q_{b,e}$ and $Q_{b,s}$. The ratio $rQes = Q_{b,e}/Q_{b,s}$ is on average generally close to 1 or slightly above 1 (Figure S1).

Above a critical discharge Q_c of about $1 \text{ m}^3/\text{s}$, bedload transport in the Erlenbach becomes approximately linearly dependent on discharge (Rickenmann, 2016, Figure 3.5 therein). Above this critical discharge begins approximately the higher intensity transport region of Equations A4a and A4b, and the corresponding shear stress ratio is about $\tau_{D50}^*/\tau_{RD50}^* = 1.28$ for $Q_c = 1 \text{ m}^3/\text{s}$. To examine whether the relations between characteristic variables are different for events with $Q_{max} < Q_c$ and for those with $Q_{max} > Q_c$, the correlation matrices were also determined separately for these two cases (Figure 4). For the subgroup with $Q_{max} < 1 \text{ m}^3/\text{s}$, there are fairly good correlations between either $Q_{b,e}$ or $Q_{b,s}$ and variables characterizing the flood strength (Q_{max} , MQ_{bt}), whereas such correlations are largely absent for the subgroup with $Q_{max} > 1 \text{ m}^3/\text{s}$. Contrary, the correlations between either $Q_{b,e}$ or $Q_{b,s}$ and variables characterizing the disequilibrium ratio (E_d) are of similar strength for both subgroups.

3.2. Temporal Variability of Disequilibrium Ratio and Memory Effects

First, I determined the autocorrelation function and the corresponding correlation coefficient R for different lag times for both the disequilibrium ratio (E_d) and the threshold conditions for start ($Q_{b,s}$) and end ($Q_{b,e}$) of bedload transport during a flood event. In doing so, the time axis is expressed by chronological event number. This choice was motivated by the fact that during interevent periods, there may be some secondary effects on $Q_{b,s}$ by low flows or by hillslope processes, but there is a

Figure 4. Pierson correlation coefficients R calculated for variables characterizing the sediment-transporting flood events. The R values were calculated in the log domain for all variables. All correlations of cells with a nonwhite background are significant with p values larger than 0.05. (a) The subgroup consisting of 367 flood events (Periods A and B) with $Q_{max} < 1 \text{ m}^3/\text{s}$ and (b) the subgroup consisting of 155 flood events (Periods A and B) with $Q_{max} > 1 \text{ m}^3/\text{s}$.

generally good correlation between these variables considering only process activities during sediment transporting flood events (Figures 3 and 4). This analysis was done separately for Periods A and B (Figures 5a and 5b), because there was a gap of about 3 years (2000–2002) between the two periods with incomplete bedload transport monitoring by the impact plate system. I have similarly checked the autocorrelation functions for other variables describing bedload transport (M_{gravel} , M_{greg}) and flow strength (MQ_{bt} , Q_{max}), but no significant correlations were found. This is not surprising, since the meteorological conditions driving the hydrology (and to a considerable degree also bedload transport) are essentially stochastic if we disregard the subsurface water storage effects (relatively small storage capacity of the flysch subsoil and relatively small catchment area) and the fact that most events are due to summer thunderstorms, with more intense rainfall events concentrated during the midsummer period. The autocorrelation analysis for E_d , $Q_{b,s}$, and $Q_{b,e}$ shows significant autocorrelation coefficients R for a lag of up to about six events (E_d , $Q_{b,s}$) and even up to about 11 events ($Q_{b,e}$) for Period A and for a lag of up to about 11 events (E_d , $Q_{b,s}$) and even up to about 21 events ($Q_{b,e}$) for Period B.

As a next step, I examined how event disequilibrium ratio (E_d) and the threshold discharge $Q_{b,e}$ at the end of the event vary over time. To do so, I determined the moving average of E_d and of $Q_{b,e}$ over five events, which is within the event lag time with an autocorrelation for both Periods A and B. The moving average was calculated using log values. This analysis revealed a pattern of cycles with E_d and $Q_{b,e}$ either above or below the equilibrium line, which was defined as the median of the moving average (Figure 6). The cycles have

Figure 5. Autocorrelation coefficients R as a function of event lag for the three time series of the variables E_d , $Q_{b,s}$, and $Q_{b,e}$. The green dashed line denotes the 95% confidence interval. (a) Period A and (b) Period B.

Figure 6. Temporal variation of E_d (log) and $Q_{b,e}$ (log) values representing a moving mean over five events. The horizontal dashed lines indicate the median values of E_d and $Q_{b,e}$, separating the above and below equilibrium periods. (a) Period A, the vertical dashed line marks the occurrence of one exceptional event (14 July 1995). (b) Period B, the vertical dashed lines mark the occurrence of two exceptional events (20 June 2007 and 1 August 2010). (Note the different E_d scales on the left ordinate for the two periods.)

different durations, they could thus be called nonperiodic. The two cycles appear to vary in an opposite way, that is, the larger the E_d , the smaller is $Q_{b,e}$ and vice versa. This suggests that there may be a feedback mechanism regulating the fluctuations around a mean value of both variables.

Based on normalized time series of E_d and of $Q_{b,e}$, each centered on its median value, I estimated the mean number of events for an average cycle (Figures 7a and 7b). This estimate is based on the number of crossing points of the two normalized time series. A Monte-Carlo simulation generating a random time sequence of the observed pairs of E_d and $Q_{b,e}$ demonstrates that (i) the number of crossing points for the observed pairs with the true chronological order (Figures 7a and 7b) is different from the number of crossing points determined for all random time sequences (Figures 7c and 7d) and (ii) there is a systematic difference in the number of crossing points between Periods A and B, which is likely due to larger maximum observed E_d values for Period B than for Period A. For an average full cycle, the mean number of events is 20.4 for Period A (number of crossing points = 28) and 26.2 for Period B (number of crossing points = 18). As for the original time series, for the random time sequences of pairs of E_d and $Q_{b,e}$ (log) values, first as moving mean over five events was calculated; then the same procedure was applied as for the real data with the original chronological order (Figures 7a and 7b), and the number of crossing points was determined (Figures 7c and 7d).

The temporal variabilities of the ratio $rQes$ and of E_d were also compared with each other, using moving average log values over five events (Figure 8). For Period B, there is a predominantly inverse correlation between the two variables. During periods with higher disequilibrium ratio, the ratio $rQes$ tends to be below the equilibrium line, implying a smaller discharge threshold for transport at the end than at the start of an event; inversely, during periods with lower disequilibrium ratio, the ratio $rQes$ tends to be above the equilibrium line, implying a higher transport threshold at the end than at the start of an event. However, the above tendency is much less pronounced for Period A, for which the inverse correlation between the two variables is visible for only about 60% of the time (events). This distinction between the two periods is also illustrated by the cross-correlation coefficients between $rQes$ and E_d , using moving average log values over five events, which are larger for Period B than for Period A (Figure S2). For Period B, the peak correlation is observed for a positive lag of about 10 events for E_d , indicating a delayed effect on $rQes$.

Figure 7. Determination of a mean cycle duration. (a and b) The time series of E_d and of $Q_{b,e}$ (log) values were first normalized to vary between 0 and 1, and then they were shifted by its median value to be centered on the zero line of the ordinate (dashed black line). These transformed (event) time series are labeled $E_{d,t}$ and $Q_{b,e,t}$, and the horizontal black line corresponds to the median values of both series, separating the above and below equilibrium periods. Based on these transformed time series, the mean number of events per full cycle was estimated by counting the number of crossing points (NCP) defined by the intersection of the two time series. The vertical dashed lines mark the occurrence of the exceptional events (as in Figure 6). (c) and (d) are histograms of NCP, based on 5,000 Monte-Carlo simulations (see text for more details), and the green vertical dashed line represents the NCP for the real data, and this NCP is clearly smaller than the NCPs for the 5,000 random time sequences. (a and c) Period A. (b and d) Period B.

It is noted that from 1986 to 2016, one of the three largest sediment-transporting flood events occurred during Period A and two during Period B (Table 2). The cumulative bedload volume over time or events indicates that about 33% more bedload was moved during Period B than during Period A (Table 2). Thus, the effects of flood events on altering the streambed characteristics may have been larger and longer lasting on average during Period B than during Period A. This is also reflected in the cross-correlations between key variables characterizing the bedload transport events (Figure 9). While the cross-correlation coefficients R between $Q_{b,e}$ and MQ_{bt} are of similar magnitude for both periods, the cross-correlations between $Q_{b,e}$ or $Q_{b,s}$ and E_d are larger for Period B than for Period A.

3.3. Event Magnitude, Channel Characteristics, and Sediment Availability

The occurrence of the three largest flood events with a peak discharge of about $10 \text{ m}^3/\text{s}$ or more (Table 2; called “exceptional” events) caused a step destruction and later reformation (Turowski et al., 2013). This resulted in an increased sediment availability afterward, which corresponds to long time periods with relatively high disequilibrium ratios E_d (Table 2 and Figures 6 and 8). From repeated longitudinal streambed profile surveys partly reported by Turowski et al. (2013), we have estimated the cumulative erosion or deposition volume along the 550 m long Erlenbach study reach upstream of the measuring site (Figure 10). For both periods and the two cases either including exceptional events or not, this indicates that the absolute maximum of deposited or eroded sediment volume was of a similar order as the sediment output during a given period (Table 3). Observed bedload mass in Table 3 is expressed as bedload transport volume V_{sed} , including pore volume and fine material, and refers either to large events or to defined time periods; thus, the bedload values are more directly comparable with eroded or deposited volumes

Figure 8. Temporal variation of E_d (log) and $rQes$ (log) values representing a moving mean over five events. The horizontal dashed lines indicate the median values of E_d and $rQes$, separating the above and below disequilibrium periods. The horizontal solid red line indicates $rQes = 1$. (a) Period A, the vertical dashed line marks the occurrence of one exceptional event (14 July 1995). (b) Period B, the vertical dashed lines mark the occurrence of two exceptional events (20 June 2007 and 1 August 2010). (Note the different E_d scales on the left ordinate for the two periods.)

indicated in Figure 10 and with estimated bedload volumes associated with a full (mean) transport cycle. A mean bedload transport volume V_{sed} per transport cycle was estimated based on two different assumptions (Table 3).

Apart from the three exceptional flood events reported in Table 2, we did not note any substantial destruction of steps in the Erlenbach since 1986 along the 550 m long study reach. This is in agreement with local erosion or deposition reported to be mostly smaller than about 0.7 m along the study reach by Turowski et al. (2013) for periods excluding the three exceptional events. For such periods, the absolute maximum of deposited or eroded sediment volume was on average less than about 300 m³/year (Figure 10), indicating smaller changes in storage volumes than during and after the occurrence of the exceptional events. It may also be noted from Figure 10 that after the exceptional event of 1 August 2010, net erosion predominated along the study reach combined with a substantial sediment output. This may be a reason for the shorter sustained period of high disequilibrium ratio after this event, as compared to the effect of the event of 14 July 1995 which was of similar magnitude (Table 2 and Figure 6).

Field observations on bedload particle displacements with tagged grains in the Erlenbach reported by Schneider et al. (2014) were used to estimate a mean active layer depth for Periods A and B. For eight events with Q_{max} values from 1 to 5 m³/s and event bedload volumes of less than 200 m³, they reported mean

Table 2
Characteristic Disequilibrium Periods Following the Three Largest Flood Events

Period	Date of exceptional flood event	Event no.	Q_{max} (m ³ /s)	MQ_{bt} (10 ³ kg)	E_d	End date of follow-up period	Number of events in follow-up period	Events above equilibrium E_d (%)	Events below equilibrium $Q_{b,e}$ (%)
A	1995-07-14	174	9.8	428	3.8	1998-09-11	78	79	82
B	2007-06-20	34	15.6	909	10.7	2010-06-19	67	87	99
B	2010-08-01	115	10.9	568	7.1	2011-07-24	37	67	85
B	2010-08-01					2013-08-27	72	40	60

Figure 9. Cross-correlation coefficients R between key variables characterizing the bedload transport events (all values represent individual events). The green dashed line denotes the 95% confidence interval. (a) Period A and (b) Period B.

transport distances from about 30 to 190 m and mean active layer depths h_a from about 0.1 to 0.23 m. Here I compare these characteristic values with conditions for typical full (mean) transport cycles for Periods A and B. The V_{sed} values for a full transport cycle given in Table 3 are about twice as large as the V_{sed} values of the above cited (medium-sized) single events from Schneider et al. (2014). Thus, it is likely that the mean exchange bedload layer thickness during a transport cycle did not clearly exceed the above reported active layer depths h_a , which are of the order of the D_{50} to D_{84} grain sizes of the bed surface material. This observation is in agreement with other field studies in similar streams reporting a largely stable coarse surface layer during average or slightly above average peak flows (Adenlof & Wohl, 1994; Church, 2010; Lenzi et al., 2004). However, for Period B, the mean V_{sed} values for a full transport cycle given in Table 3 range from about 680 to 430 m³ (with or without exceptional events). These values are roughly comparable to the V_{sed} of 519 m³ for the exceptional event of 14 July 1995 (Table 3), and for the survey period including this event, Schneider et al. (2014) estimated an active layer depth h_a of 0.57 m. Thus, it may be assumed that the bedload transported during a full cycle in Period B had built up a larger active layer, which required more time (or events) to be reworked (and partly removed) by later events. Substantial changes to the bed surface after large floods or sediment inputs were also observed in other field studies in mountain streams (Gintz et al., 1996; Lenzi et al., 2004; Reid et al., 2019).

Figure 10. Constructed sediment budget along 550 m long study reach for different time periods. Storage volume (S) is cumulative erosion (negative value) or deposition (positive value) volume (including pore space) estimated from repeated longitudinal streambed profile surveys (from Leuzinger, 2011, unpublished master thesis), sediment output (O) is determined from the impact plate measurements, and sediment input (I) is estimated as $I = S + O$. The stars in parentheses indicate the number of exceptional events within a given survey period. The data used to produce this figure and some more details are given in Table S2.

Table 3

Bedload Transport Volume per Transport Cycle Estimated From a Mean Event Volume of All Events (a) or From a Mean Value Disregarding the Three Exceptional Events From Table 2 (b)

	Period A	Period B
Mean number of events per full cycle	20.4	26.2
Mean number of events per year	22.0	16.9
Bedload volume for entire period (m ³) (a)	4,493	6,084
Total number of events per period	286	236
Mean volume per event for period (m ³) (a)	15.7	25.8
Mean volume per full cycle (m ³) from (a)	321	675
Bedload volume of exceptional events (b)	519	2,243
Mean volume per full cycle (m ³) without (b)	284	430

Most sediment-transporting flood events occurred during the summer half year, with more frequent events from June through August when the typical triggering weather conditions with convective thunderstorms are most likely. This is reflected in the seasonal peak of the $Q_{\max}$ values of the events occurring during these same months, and a similar trend is observed for the variables directly depending on the magnitude of the flood events. Contrary, there is no similar trend for a seasonal variation of E_d , whereas for $Q_{b,s}$ and for $Q_{b,e}$, a similar seasonal trend is visible only for Period A but is largely absent for Period B. Comparing also the mean number of events per full cycle with the mean number of events per year (Table 3) indicates that the mean cycle duration for Period A is likely related to the seasonal fluctuation of the hydraulic forcing, whereas this is not the case for Period B.

3.4. Prediction of Disequilibrium Ratio Using Memory Effect and Hydraulic Forcing

Given the autocorrelation of the disequilibrium ratio and threshold transport conditions and the cross-correlation between the two variables (Figures 5 and 8), I examined whether bedload-transport prediction can be improved when using information of prior events. A correlation matrix was determined for the variables M_{gravel} , E_d , M_{greg} , and $Q_{b,s}$ of a given event and variables characterizing the prior event (subscript _p) such as $Q_{b,e,p}$, $Q_{\max,p}$, $MQ_{bt,p}$, and $E_{d,p}$ (Figure S3). There is a significant correlation between E_d , M_{greg} , and $Q_{b,s}$ and $Q_{b,e,p}$, $E_{d,p}$, and $MQ_{bt,p}$. Using (stepwise) regression analysis, the following power law equations were determined to predict the event bedload mass M_{g_pred1} and M_{g_pred2} :

$$M_{g_pred1} = 7.08 MQ_{bt}^{0.76} E_{d,p}^{0.38} Q_{b,e,p}^{-0.71}, \quad (4)$$

$$M_{g_pred2} = 5.07 MQ_{bt}^{0.75} Q_{b,e,p}^{-1.36}. \quad (5)$$

The exponent of the factor MQ_{bt} in Equations 4 and 5 is not too dissimilar to the corresponding exponent in Equation 2. The performance of the two equations is shown in Figure 11, and the associated statistics of all three predictive equations for event bedload masses are given in Table 4. M_{cal} is the notation for the predicted bedload mass using either Equation 2, 4, or 5. The variability of the ratio $M_{\text{gravel}}/M_{\text{cal}}$ is a measure of the predictive quality of the three equations. For Equation 2, it is expressed by E_d (Equation 3). The density distribution of this ratio for the three predictive equations is further illustrated in Figure S4. Accounting for

Figure 11. Comparison of observed bedload mass M_{gravel} for each flood event and calculated bedload mass M_{cal} . The black line is a power law regression between the two variables, and the red dotted line is the 1:1 line. (a) $M_{\text{cal}} = M_{g_pred1}$, using Equation 4. (b) $M_{\text{cal}} = M_{g_pred2}$, using Equation 5.

Table 4
Statistics of All Three Predictive Equations M_{cal} for Event Bedload Masses Compared to the Observed Event Bedload Masses M_{gravel} Determined for the Same Total of 520 Events

Equation no. for M_{cal}	Pierson correlation coefficient R (using log values)	Standard deviation of $\log(M_{gravel}/M_{cal})$
(3)	0.839	3.30
(6)	0.898	2.32
(7)	0.884	2.45

the memory effect improves the prediction of bedload transport for the next event, as is also shown in Table 4. Interestingly, including the bed state after the previous event ($Q_{b,e,p}$) alone is already a good indicator to improve the prediction, while additional inclusion of the prior ($E_{d,p}$) results in a limited further improved prediction.

4. Discussion

4.1. Influence of Runoff and Disequilibrium Ratio on Threshold Conditions

Masteller et al. (2019) investigated the effect of runoff conditions on the threshold for the start of bedload transport in the Erlenbach for the Years 1987 to 1999 and 2007 to 2012. They explored the temporal evolution of the threshold shear stress with time for each year separately. Significant trends of threshold shear stress and time with a positive correlation with $R > 0.6$ were found for the following seven years: 1987–1989, 1996–1997, 2008, and 2012. For most of these years, we found a relatively strong correlation between threshold discharge $Q_{b,s}$ and event-integrated bed shear stress MQ_{bt} (Figure S5). However, for the Years 2009–2011, such a correlation is much weaker (Figure S5b).

Regarding interevent flows, Masteller et al. (2019) observed a significant positive correlation between the threshold shear stress and mean interevent shear stress above 104 Pa (corresponding to $Q_o = 0.14 \text{ m}^3/\text{s}$ in their analysis), which was thought to have caused a reorganization and structuring of the bed. Analyzing event flows, they also found a significant positive correlation between the threshold shear stress (τ_c) and the peak transporting shear stress of the previous transport event ($\tau_{\max,p}$), followed by an inverse correlation between τ_c and $\tau_{\max,p}$ for shear stresses above about 340 Pa (corresponding to $Q_o = 1.38 \text{ m}^3/\text{s}$ in their analysis).

The degree of correlation in Figure 4 between E_d , $Q_{b,e}$, and $Q_{b,s}$ indicates a possibly important influence of flood variables on the threshold discharges only for the subgroup with $Q_{\max} < 1 \text{ m}^3/\text{s}$, that is, for those events where the streambed conditions are likely to be less strongly affected by moving bedload. Thus, our findings confirm the results of Masteller et al. (2019) that low to moderate runoff can have a stabilizing and armoring effect on the streambed. On the contrary, we found a negative correlation between $Q_{b,e}$ and $Q_{\max}$ for events with $Q_{\max} > 1 \text{ m}^3/\text{s}$ (Figure 4). Our results indicate a possible important influence of a larger disequilibrium ratio E_d on reducing the threshold discharges $Q_{b,e}$ and $Q_{b,s}$ (or the other way round) for all flow conditions (Figure 3). Strictly speaking, a causal effect can be expected only from E_d on $Q_{b,e}$ and from $Q_{b,s}$ on E_d for a given event and not the other way round. However, as there is substantial autocorrelation and cross-correlation for all these variables (s. section 3.2), this clear distinction is no longer valid. Our data also indicate a trend that $Q_{b,e}$ tended to be larger than $Q_{b,s}$ for most of the time during Period A and for the beginning and end of Period B. On the contrary, for a duration of about 4 years after the extreme events of 20 June 2007 and 1 August 2010 until 15 August 2011, $Q_{b,e}$ tended to be similar as $Q_{b,s}$ (Figure 8). This latter trend includes roughly half of all events of Period B and may be a result of increased sediment availability on the bed for this time period (Figure 10). Excluding this 4 year period, events with low and moderate bedload-transport transporting flows tended to result in $Q_{b,e} \approx 1.2 Q_{b,s}$ on average.

A final remark concerns the fact that the threshold discharge at the start of an event $Q_{b,s}$ can be expected to have some similarity with the $Q_{b,e,p}$ of the previous event, which is also confirmed from the analysis including prior-event information (Figure S3). In an earlier study, Turowski et al. (2011) showed a rather strong correlation between $Q_{b,s}$ and $Q_{b,e,p}$ for some Austrian glacial streams, while there was much more scatter between the two variables for the Erlenbach. This variability may have been due partly to hillslope processes affecting the bed surface texture and morphology (Turowski et al., 2011) and partly to interevent flows stabilizing the bed surface (Masteller et al., 2019). A detailed discussion about these effects on the threshold discharge is given in the two cited publications.

4.2. Cycle Duration and Sediment Supply or Availability

As it is evident from Table 3, the mean (full) cycle duration is about 20 events for Period A and about 26 events for Period B, which translates to about 0.9 years for Period A and about 1.6 years for Period B.

Figure 12. Conceptual sketch illustrating the main differences in the patterns of disequilibrium transport cycles between Periods A and B. The scaling of the two graphs is approximate and represents the conditions in the Erlenbach stream. The maxima and minima values indicated for the two periods were determined as the mean of the two largest or smallest five-event-average values (Figure 6) for the two considered variables. Note that for a given period, observed cycles had different durations, whereas the cycles shown here have the same (mean) duration.

While the range of variation of the threshold transport discharge $Q_{b,e}$ is similar for both periods, the maximum disequilibrium ratio E_d for Period B with 10.7 is clearly larger than for Period A with 3.8 (Table 2, values for a single event). Disequilibrium ratio and threshold transport discharge are inversely correlated with each other (Figure 6), and they appear to provide a feedback mechanism governing the fluctuations around an equilibrium state. The main differences in the disequilibrium transport cycle patterns between Periods A and B are illustrated in a conceptual sketch in Figure 12. The mean cycle duration is probably affected by sediment supply to or availability (Figure 10) in the reach upstream of the retention basin (measuring cross section). It may be hypothesized that a larger sediment input results in a thicker exchange layer and looser bed structures, favoring comparatively smaller threshold transport discharges. This will increase bedload transport capacity and sediment availability resulting in larger disequilibrium ratios. Larger bedload transport out of the reach will diminish the exchange layer, and discharge periods with intermediate to weak transport intensities will stabilize or armor the bed, thus increasing the threshold transport discharge.

Field studies that documented similar sediment transport cycles over long time periods are rare. Based on a 45 year data set with annual- and decadal-scale observations on sediment storage, channel morphology, and wood loading, Reid et al. (2019) analyzed the spatial and temporal organization of storage along a 3 km reach of the main channel in a previously glaciated 11 km² catchment in British Columbia, Canada. They found typical aggradation/degradation cycles over time, and they concluded that changes in sediment storage through time and sediment transfer through the system are not simply a function of flow conditions but also of variable sediment supply conditions. For their study reach, small-scale aggradation/degradation cycles for stored sediment typically varied between a few and about 12 years. These long-term observations are in qualitative agreement with the interpretation of our long-term sediment transport measurements at the Erlenbach.

Elgueta-Astaburuaga et al. (2018) analyzed the effect of sediment supply on sediment mobility for a poorly sorted experimental bed, and they suggested that large sediment pulses may increase the strength and persistence of autocorrelation in bedload rate time series. Elgueta-Astaburuaga and Hassan (2019) further studied the adjustment of a gravel bed during cycles of sediment storage in a flume experiment. They found that large cycles were caused by significant changes in external sediment supply and small cycles were the result of local changes in bed conditions. Though we do not have information on local changes in bed conditions at the Erlenbach, our observations confirm the important role of sediment supply. We observed a similar effect at the Erlenbach, when comparing the autocorrelation (Figure 5) and cross-correlation (Figure 9) of some key variables between Periods A and B, the latter being characterized by larger sediment supply and longer cycle duration.

In the Erlenbach, the three largest flood events with a peak discharge of about 10 m³/s or more caused a step destruction and later reformation and resulted in a considerably increased sediment availability afterward, corresponding to long time periods with relatively high disequilibrium ratios E_d (Table 2 and Figures 6 and 7). Such effects of exceptionally large flood events on bedload transport in the aftermath have been reported also for other steep mountain streams (Ashida et al., 1976; Gintz et al., 1996; Lenzi et al., 2004). The possible reasons for this increase were summarized by Turowski et al. (2009) as follows: destruction or rearrangement of bed structures reducing form roughness and increasing transport capacity; reduced imbrication and interlocking of grains; and destabilized banks and increased sediment supply from the hillslopes.

4.3. Variability Around Equilibrium Transport Conditions

For Period A, the mean transport cycle duration of the disequilibrium ratio E_d and the threshold discharge $Q_{b,e}$ (Figure 6) approximately corresponds to a (hydrological) year which is associated with typical seasonal variations of flood and bedload transport magnitude. This is not the case for Period B, for which the mean full cycle duration is approximately 1.6 years (Table 3). In this regard it is important to note that event bedload, expressed as M_{gravel} or V_{sed} , generally shows more frequent fluctuations around a mean value (Figure S6) than what is observed for the transport cycles of E_d and $Q_{b,e}$ around a mean value. This suggests that the disequilibrium ratio (E_d) and the threshold discharge ($Q_{b,e}$, $Q_{b,s}$) primarily control the fluctuations of bedload transport around equilibrium conditions, whereas sediment supply may have an effect on the duration of such cycles. The hydrology (flood magnitude) and hillslope instabilities generally have a primary effect on sediment supply, but the flood magnitude itself does not seem to directly govern the duration of a cycle.

The identification of transport cycles with fluctuating values of E_d and $Q_{b,e}$ is in qualitative agreement with the model of Johnson (2016) describing disequilibrium conditions of gravel transport. The variation of the $Q_{b,e}$ values ranges from about 0.2 to 1.0 m³/s, and according to the hydraulic calculations, this corresponds to a range of critical Shields stresses of $0.13 < \tau_{D50}^* < 0.23$. Johnson (2016) found from his experiments that aggradation is more efficient at decreasing critical Shields stress than degradation is at increasing critical Shields stress. Our observations provide support for this hypothesis if we consider the effect of the three exceptional events. For Period A, it appears that after the occurrence of the exceptional event on 14 July 1995, there was a decrease of the threshold discharge $Q_{b,e}$ over about 20 events, followed by a rather slow overall increasing trend over about 80 events, that is, until about June 1999 (Figure 6a). For Period B, similar patterns are observed after the occurrence of the exceptional event on 20 June 2007 when there was a rapid decrease of $Q_{b,e}$ over about five events, followed by an increase over about 20 events until July 2008, and after the occurrence of the exceptional event on 1 August 2010 when there was an overall decrease of $Q_{b,e}$ over about 25 events, followed by an increase over about 55 events until about July 2014 (Figure 6b).

From the channel reach observations described in section 3.3, I have hypothesized that the exceptional events, which had high disequilibrium ratios (Table 2), resulted in a larger thickness of the exchange (“active”) layer than for the periods with lower flow magnitudes, which had an estimated exchange layer thickness on the order of D_{50} to D_{84} . Protrusion upstream of boulder steps was measured at the Erlenbach by Yager et al. (2012) five times during the period of 2004 through September 2010, which included the occurrence of two exceptional events. They demonstrated that the protrusion systematically declined with greater measured relative sediment availability, and they found that protrusion can be described as a function of the time elapsed since an exceptional event. These findings support the interpretation that the sediment availability was higher during periods with high E_d values and particularly after exceptional events. Yager et al. (2012) also noted that protrusion affects bed roughness, and they measured a finer grain size distribution after the extreme event of 20 June 2007 as compared to 2004. Thus, periods with low protrusion (and high E_d values) are in qualitative agreement with periods of smaller threshold discharges, confirming our finding of an inverse relation between E_d and $Q_{b,e}$.

Chartrand et al. (2019) proposed that the local disequilibrium state of gravel-bed mountain streams may be governed by (i) the rate of local bed topography adjustment and (ii) the rate of bed sediment texture adjustment. Based on flume experiments, they suggested that when the ratio of bed shear stress to reference (threshold) shear stress varied from about 1.5 to 2.0, the dominant control on disequilibrium shifted from Element (i) to Element (ii). At the Erlenbach, this relative shear stress intensity varying from 1.5 to 2.0 corresponds to discharges Q_o ranging from 1.3 to 2.5 m³/s. Our analysis and observations at the Erlenbach suggest that during periods with lower E_d values, changes in bed texture may have been more important than changes in local bed topography, because of a generally smaller exchange layer thickness and relatively more important protrusion of larger grains; contrary, changes in local bed topography may have become relatively more important after the occurrence of the exceptional events. It is thus possible that relative shear stress intensity may not be the only criterion to determine the important elements of control on disequilibrium in the Erlenbach.

Rickenmann (2018) analyzed 6 years of continuous bedload transport measurements in two Austrian mountain streams. His study suggested the presence of quasi-stable periods of days to weeks for which a

constant critical Shields stress may be associated with quasi-stable conditions of the bed morphology upstream of the measuring cross section. These quasi-stable periods may be similar to the disequilibrium transport cycles identified for the Erlenbach. The observations from the two Austrian mountain streams (Rickenmann, 2018) further indicated that about 100 days of measurements are required in the Fischbach and about 70 days in the Ruetz, to arrive at a mean estimate of an equilibrium transport level that was within a factor of 2 in relation to the 6 year average (including about 700 days of transport activity), based on cumulative values of Q_b and Q . This mean transport level is possibly representative of equilibrium transport conditions. If cumulative values of Q_b and Q_o are similarly determined over time (i.e., event number) for the Erlenbach (Figure S7), one can estimate the number of events necessary to reduce the variability of the ratio of $\text{Sum}(M_{\text{gravel}})/\text{Sum}(Q_o)$ to within a factor of 2 in relation to the long-term average of all events per period, resulting in about 15 events required for the Period A and about 33 events required for the Period B. This corresponds to less than one full year for Period A and to about 2 years for Period B. The occurrence of the relatively rare exceptional events (with Q_{max} of $10 \text{ m}^3/\text{s}$ or more), which have a recurrence period of about 14 years at the Erlenbach, increases both the mean transport level and the observation time required to arrive at a relatively close estimate of a mean transport level.

In flume experiments, Saletti et al. (2015) observed that the variability in sediment transport can be explained as a consequence of the stochastic nature of bed stability, that high-magnitude rates play an important role in bulk sediment transport, and that some of the bursts in transport rates result from the displacement of large grains. At the Erlenbach, the maximum transported grain sizes were smaller than about 0.1 to 0.2 m for discharges up to about $1 \text{ m}^3/\text{s}$ (Rickenmann et al., 2012), whereas for the extreme event of 20 June 2007, the 10 largest grains deposited in the retention basin had b axes from 0.45 to 0.63 m, and even larger boulders (b axes from 1.0 to 1.35 m) were moved in the channel (Turowski et al., 2009). Thus, it is likely that coarser particles play an important role regarding the variability of bedload transport rates in the Erlenbach, particularly for flows with discharges larger than about $1 \text{ m}^3/\text{s}$.

5. Conclusions

For the Erlenbach, a Swiss mountain stream, indirect bedload transport measurements with impact plates were analyzed for more than 500 flood events that occurred during a 27 year long observation period. Different impact sensors were used, and therefore, two different periods were defined, namely, Period A from 1986 to 1999 and Period B from 2002 to 2016. Hydraulic and bedload transport calculations were used to determine a disequilibrium ratio for each event. Apart from continuous records of bedload transport and discharge, some additional measurements were available from repeated surveys of the longitudinal bed profile along a 550 m study reach upstream of the Erlenbach bedload measuring site. Moreover, some information on mean particle transport distances and virtual active layer depths from earlier particle tracer studies was used.

The measured bedload mass per event was divided by the bedload-transport capacity integrated over a flood event to define a disequilibrium ratio. Larger disequilibrium ratios corresponded to lower threshold transport discharges at the end of a flood event and vice versa. Both variables show a significant autocorrelation over at least five events and also a significant cross-correlation with each other. As a result, disequilibrium ratio and threshold discharge varied in a cyclic (but nonperiodic) behavior around an equilibrium ratio defined by its median value. Thus, a memory effect was evident both regarding the disequilibrium ratio and the bed state as reflected by threshold transport conditions. The mean duration of a disequilibrium transport cycle was different for Periods A and B, probably depending on sediment supply/availability (magnitude of previous events). This interpretation is also based on the occurrence of three exceptional flood events which resulted in the destruction of many steps along the study reach and a rather rapid reformation of steps by follow-up floods. Using these findings together with information from earlier studies in the Erlenbach, it is suggested that the mean thickness of the exchange (active) layer on the bed surface may have been about one grain size (D_{50} to D_{84}) during Period A, whereas during Period B, the mean exchange layer thickness may have been about twice the size of the bed surface D_{84} . If the memory effect of the bed state, that is, the threshold discharge after the previous event, is accounted for, this improved the prediction of bedload transport for the next event.

Figure A1. Comparison of observed bedload mass M_{gravel} for each flood event and calculated bedload mass MQ_{bt} , using Equations A4a and A4b and integrating over the duration of the flood event. The black line is a power law regression between the two variables (see also Equation 2), and the red dotted line is the 1:1 line.

Appendix A: Details of Hydraulic and Bedload-Transport Calculations

The lowermost channel reach upstream of the sediment retention basin (with a natural bed) has a trapezoidal cross section, including partially engineered banks protected by a riprap construction (Figure 1c). Several cross sections were surveyed with a total station; the average bottom width is 4.1 m, and the banks have a lateral slope of 1.5:1 (Figure 1c). The hydraulic calculations were made with an equation given in Nitsche et al., (2012, Figure 5d therein) using two dimensionless variables U^{**} and q^{**} for the mean flow velocity U and the unit discharge q , respectively:

$$U^{**} = 1.49 q^{**0.6}, \quad (\text{A1})$$

$$U^{**} = U / \sqrt{gSD_{84}}, \quad (\text{A2})$$

$$q^{**} = q / \sqrt{gSD_{84}^3}, \quad (\text{A3})$$

where g is the gravitational acceleration, S is the channel bed slope, and D_{84} is the grain size of the bed surface for which 84% of the particles are finer. Equation A1 is based on dye tracer measurements made in the lowermost reach of the Erlenbach. The unit discharge q was determined for a mean width for a given flow depth in the trapezoidal cross section, and bank resistance was accounted for by reducing q with the ratio of

the hydraulic radius r_h to the flow depth h . This required an iterative calculation procedure, using the measured discharge Q_0 along with the trapezoidal cross section.

Finally, bedload-transport calculations were made with an equation given in Schneider et al. (2015, Equation 13 therein), which represents a modified form of the Wilcock and Crowe (2003) equation:

$$W_{\text{tot}}^* = 0.002 \left(\frac{\tau_{D50}^*}{\tau_{rD50}^*} \right)^{16.1} \quad \text{for } \frac{\tau_{D50}^*}{\tau_{rD50}^*} < 1.143 \text{ and } D > 4 \text{ mm}, \quad (\text{A4a})$$

$$W_{\text{tot}}^* = 14 \left(1 - \frac{0.85}{(\tau_{D50}^* / \tau_{rD50}^*)^{0.7}} \right)^{4.5} \quad \text{for } \frac{\tau_{D50}^*}{\tau_{rD50}^*} \geq 1.143 \text{ and } D > 4 \text{ mm}, \quad (\text{A4b})$$

where the dimensionless transport rate W^* is defined as

$$W^* = Rgq_b / u^{*3}, \quad (\text{A5})$$

where $R = \rho_s / \rho - 1$ is the relative sediment density (with ρ_s = sediment density and ρ = water density), q_b is the volumetric bedload transport rate per unit width, $u^* = (\tau / \rho)^{0.5}$ is the shear velocity, and $\tau = g\phi r_h S$ is the bed shear stress. $\tau_{D50}^* = r_h S / (RD_{50})$ is the dimensionless bed shear stress with regard to the characteristic grain size D_{50} , and τ_{rD50}^* is the dimensionless reference bed shear stress with regard to the characteristic grain size D_{50} of the bed surface. Equations A4a and A4b were developed for calculation of total transport rates (not fractional transport rates). The threshold between low- and high-intensity transports in Equations A4a and A4b has been corrected to $\tau_{D50}^* / \tau_{rD50}^* = 1.143$, as compared to the value of 1.2 given in Schneider et al. (2015). τ_{rD50}^* is calculated as a function of the bed slope (Schneider et al., 2015, Equation 10 therein):

$$\tau_{rD50}^* = 0.56 S^{0.5} = 0.18. \quad (\text{A6})$$

Using Equations A4a and A4b and integrating the transport rate for the entire channel width over the duration of the flood event resulted in the “predicted” bedload mass MQ_{bt} (in kg) for each flood event. It is noted that for the development of Equations A4a and A4b, Schneider et al. (2015) used bedload transport

measurements from 14 mountain streams, including samples from the Erlenbach that had been obtained with the moving basket system (Rickenmann et al., 2012).

The relation between the observed event bedload mass M_{gravel} and the calculated event bedload mass MQ_{bt} has a correlation coefficient $R = 0.836$ (using log values), and it can be approximated by a power law regression between the two variables, as is illustrated in Figure A1.

Data Availability Statement

The bedload transport and runoff data integrated over the flood event duration are available online (<https://www.envidat.ch/#/metadata/sediment-transport-observations-in-swiss-mountain-streams>).

List of Variables

E_d	Disequilibrium ratio (Equation 3)
$E_{d,p}$	E_d value of prior event
h_a	Active layer depth (thickness of sediment exchange layer)
M_{cal}	Calculated bedload mass, using M_{greg} (Equation 2), $M_{\text{g_pred1}}$ (Equation 4), or $M_{\text{g_pred2}}$ (Equation 5)
M_{gravel}	Event-integrated measured bedload mass (Equation 1)
M_{greg}	Improved estimate of the event bedload mass (Equation 2), with a correction of the trend for underestimating the M_{gravel} values for smaller events when using the MQ_{bt} predictions
MQ_{bt}	Predicted bedload mass, based on the event-integrated bedload transport capacity formula (Equations A4a and A4b)
$Q_{b,e}$	Discharge threshold at the end of an event
$Q_{b,e,p}$	$Q_{b,e}$ value of prior event
$Q_{b,s}$	Discharge threshold at the start of an event
Q_{max}	Peak discharge of an event
Q_o	Water discharge measured at the upper-flow gauging station
Q_u	Water discharge measured at the lower-flow gauging station
rQ_{es}	Ratio $Q_{b,e}/Q_{b,s}$
V_{sed}	Observed bedload transport volume, including pore volume and fine material

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