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Key Points:
- A 19-year record of sediment transport reveals strong seasonal trends in the onset of sediment motion
- Threshold for sediment motion exhibits strong dependence on the magnitude of past flows
- Low to intermediate prior flows lead to increased channel stability, whereas higher magnitude flows can destabilize the channel bed

Supporting Information:
- Supporting Information S1
- Data Set S1
- Data Set S2

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History-Dependent Threshold for Motion Revealed by Continuous Bedload Transport Measurements in a Steep Mountain Stream

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Abstract To explore the causes of history-dependent sediment transport in rivers, we use a 19-year record of coarse sediment transport from a steep channel in Switzerland. We observe a strong dependence of the threshold for sediment motion (τ_c) on the magnitude of previous flows for prior shear stresses ranging from 104 to 340 Pa, resulting in seasonally increasing τ_c for 10 of 19 years. This stabilization occurs with and without measureable bedload transport, suggesting that small-scale riverbed rearrangement increases τ_c. Following large transport events (>340 Pa), this history dependence is disrupted. Bedload tracers suggest that significant reorganization of the bed erases memory of previous flows. We suggest that the magnitude of past flows controls the organization of the bed, which then modifies τ_c, paralleling the evolution of granular media under shear. Our results support the use of a state function to better predict variability in bedload sediment transport rates.

Plain Language Summary Fluvial bedload transport governs river channel evolution and has important implications for river restoration efforts, aquatic habitats, and water quality. This fundamental sediment transport process is typically modeled independently of prior flow history, despite previous observations suggesting its influence. We evaluate the role of past flows in moderating channel stability with a unique, long-term data set from a Swiss mountain stream. We find that the onset of bedload transport has memory of past flow magnitude, where small to intermediate past flows build channel stability. High-magnitude flows disrupt this memory and can destabilize the channel bed. These results challenge the assumptions of widely used bedload transport models and suggest that predictions of erosion in river channels may be improved by accounting for the effects of prior flows.

1. Introduction

Virtually all problems related to fluvial morphodynamics, from understanding alluvial channel geometry (Parker, 1979) to modeling bedrock incision (Sklar & Dietrich, 2004) require some prediction of bedload transport rates. Despite over a century of work related to this process, predicting bedload flux represents a fundamental difficulty in geomorphology due to the complex interactions of individual sediment grains and turbulent, open channel flow (Chen & Stone, 2008; Frey & Church, 2009; Jerolmack & Paola, 2010; Wilcock & Crowe, 2003). A consequence of this interaction is the well-documented variation in the shear stress at the onset of sediment motion (Buffington & Montgomery, 1997), a quantity that underpins most predictive models for bedload transport (Engelund & Fredsøe, 1975; Fernandez Luque & Van Beek, 2010; Meyer-Peter & Müller, 1948; Wong & Parker, 2006).

In the most commonly used formulas, bedload flux is a function of dimensionless shear stress, τ^*, in excess of a defined critical value, τ^*_c (e.g., Wong & Parker, 2006). This dimensionless shear stress, commonly termed critical Shields stress, or τ^*_c, describes the flow conditions at the onset of measurable sediment transport. Typically, τ^*_c is assigned a single value for the median grain size of the channel bed surface. While bedload transport models with a single τ^*_c have significant utility owing to the minimal number of input parameters required, they offer limited physical insight and can, at times, result in up to 2 orders of magnitude of error when compared to direct transport measurements (Bathurst, 2007; Gomez & Church, 1989; Lenzi et al., 2006; Recking et al., 2012; Scheingross et al., 2013; Schneider et al., 2015). This mismatch implies that these models may not adequately capture bedload transport dynamics, especially in steep streams (Lenzi et al., 2006; Scheingross et al., 2013) and for flows very close to τ^*_c (Gomez et al., 1989; Recking et al., 2012). This disagreement represents a significant model shortcoming, as a large portion of fluvial sediment transport is thought to occur at near-threshold conditions (e.g., Parker, 1979). Thus, an accurate description of τ^*_c is one important factor in improving the prediction power of sediment transport models.

The threshold for motion varies with surface grain size distribution (Chen & Stone, 2008; Moog & Whiting, 1998), channel slope (Lamb et al., 2008), sediment supply (Dietrich et al., 1989), degree of bed compaction (Charru et al., 2004; Houssais et al., 2015), and flood intermittency (Haynes & Pender, 2007; Masteller & Finnegan, 2017; Ockelford & Haynes, 2013). Thus, the development of a more physically realistic treatment of τ^*_c represents a key knowledge gap in fluvial geomorphology. To account for this complexity, both probabilistic (e.g., Einstein, 1950; Furbish et al., 2012; Monsalve et al., 2016) and state-dependent models (Johnson, 2016) for sediment transport have been proposed. In the latter model, τ^*_c evolves through time by integrating the effects of past flows.

Reid et al. (1985) were, to our knowledge, the first to suggest that τ^*_c depends on prior flow history. They found that bedload transport rates in a perennial gravel stream were reduced during more isolated transport events compared to those that were closely spaced but did not consider variability in flow magnitude. They hypothesized that during sustained interevent flow, local sediment rearrangement occurs, increasing particle interlocking and τ^*_c. Subsequently, flume experiments have also demonstrated that bedload transport is highly sensitive to the magnitude and duration of prior interevent flows, even in the absence of measurable sediment transport (Haynes & Pender, 2007; Mao, 2018; Masteller & Finnegan, 2017; Monteith & Pender, 2005; Ockelford & Haynes, 2013; Paphitis & Collins, 2005).

Sediment mobility is also influenced by high-magnitude, sediment transporting flows. Exceptionally high discharge events are thought to reduce τ^*_c by disrupting particle interlocking through boulder mobilization and the breaking of channel steps (Turowski et al., 2009; Yager et al., 2012). This disruption in particle interlocking results in elevated bedload transport rates (Lamarre et al., 2008; Oldmeadow & Church, 2006), which can persist across numerous transport events (Yager et al., 2012). Flume experiments by Masteller and Finnegan (2017) found an increase in the number of highly mobile, high protruding grains in response to sediment transporting flows, suggesting a concurrent decrease in τ^*_c. These observations suggest that gravel beds integrate the effects of both interevent and transporting flows by reorganizing surface grains, leading to a history-dependent τ^*_c.

The sensitivity of τ^*_c to flow history has yet to be extensively explored in a natural channel. Compared to idealized flume experiments, natural channels can have complex channel structure, variable hydrograph shapes, discharge magnitudes, and event timing. Steep mountain streams are arguably the most complex, in that they are typically composed of persistent channel steps, can exhibit strong hillslope-channel coupling, and have wide grain size distributions that typically include in-channel boulders (Montgomery & Buffington, 1997). Much of the existing work on stress history effects has avoided these complexities by focusing on low gradient systems (Haynes & Pender, 2007; Monteith & Pender, 2005; Ockelford & Haynes, 2013; Paphitis & Collins, 2005; Reid et al., 1985). Thus, examination of whether or not stress history effects persist in these higher gradient channels may illuminate the universality and overall importance of stress history effects across fluvial systems as a whole.

2. A Systematic Exploration of Flow History Effects in a Natural Channel

The Erlenbach torrent is a small step-pool channel with a drainage area of 0.74 km^2 and an average gradient of 17% (Rickenmann, 2001). It is one of the most well-instrumented rivers in the world, with 30 years of high-resolution sediment transport, hydrologic, and meteorologic data (Rickenmann & McArdell, 2007). Sediment transport at the Erlenbach is driven by summer thunderstorms, leading to variable spacing between events (Schuerch et al., 2006). Bedload sensors have monitored sediment impacts at 1-min resolution since 1986 (see Rickenmann & McArdell, 2007; Rickenmann et al., 2012, for description of instrumentation). Continuous measurements of channel discharge are recorded at 10-min intervals (until 2009) and at 1-min intervals (from 2009) 30 m upstream of the impact sensors (Rickenmann & McArdell, 2007).

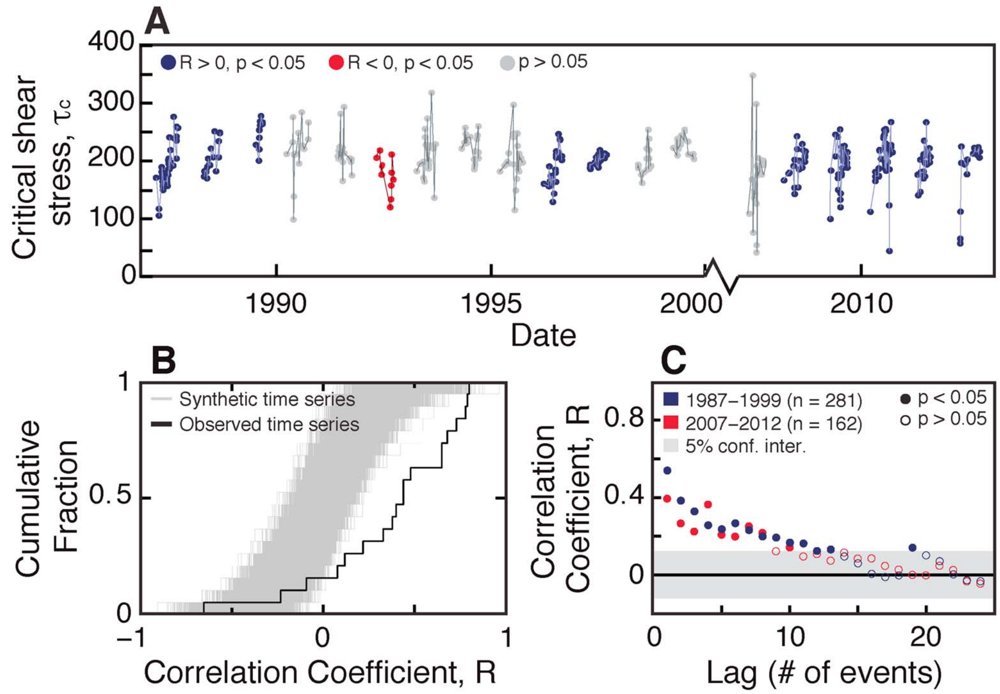

Figure 1. (a) Time series of critical shear stress (τ_c) at the Erlenbach. Years with significant, positive correlation between the critical τ_c and time ($p < 0.05$) are shown in blue, significant and negative in red, and without significant correlation ($p > 0.05$) are in gray. (b) Distribution of R values from Erlenbach time series (black) and 10,000 synthetic distributions (gray). (c) Autocorrelation of τ_c time series for years 1987–1999 (blue) and 2007–2012 (red). Filled circles indicate significant correlation ($p < 0.05$), and open circles represent no correlation ($p > 0.05$).

We extend a record of the start of bedload transport at this site, first compiled by Turowski et al. (2011) and use the data set to explore the patterns and causes of variability in the onset of coarse sediment transport, which was not addressed by the previous study. The onset of motion is defined by a nonzero number of impulses as registered by the geophones, cross checked with the timing of increases in channel discharge. We transform channel discharge measurements to boundary shear stress, τ, following the rating curve developed by Yager (2006). Our record spans February to September of each year, as discharge measurements may be unreliable during icy conditions, for 19 years, from 1987 to 1999 and 2007 to 2012 (443 transport events). There is a data gap from 1999 to 2002 during the replacement of the piezoelectric impact sensors with geophone impact plates. Years from 2002 to 2006 feature less than 10 transport events each and have been removed to avoid small sample sizes.

We utilize this unique, 19-year-long record of bedload transport at the Erlenbach to explore the variation and evolution of τ_c. This transport record allows, to our knowledge, for the most extensive and systematic examination of the dependence of the threshold for motion on flow history to date. Across the time series, periods of measured sediment transport comprise only 1.2% of the full record, whereas interevent periods without measurable bedload transport occupy the remaining 98.8%. We examine the dependence of τ_c on both interevent flows and transporting flows in order to evaluate the hypothesis that the threshold for motion is sensitive to flow history. We focus on the dimensional quantity, τ_c, rather than dimensionless Shields stress as it avoids assumptions on the size of the transported sediment.

3. Results

3.1. Seasonal Variations in the Onset of Motion

We find that for 11 of the 19 years on record (58%), variations in τ_c are significantly correlated with time by calculating the Pearson correlation coefficient between τ_c and time for each record year (Figure 1a and Table 1). Ten of these years show significant increases in τ_c from February to September (mean increase

10.1029/2018GL081325

Table 1
Correlation Coefficient Analysis of τ_c by Year

Year	Number of transport events	R, correlation coefficient	p value
1987	*46*	*0.8004*	*2.46E−11*
1988	*20*	*0.7974*	*2.55E−05*
1989	*15*	*0.7820*	*5.73E−04*
1990	16	0.2194	0.414
1991	18	−0.2611	0.295
1992	*11*	*−0.6278*	*0.039*
1993	21	0.4041	0.069
1994	20	−0.0861	0.718
1995	24	0.0699	0.746
1996	*27*	*0.6776*	*1.03E−04*
1997	*22*	*0.6540*	*9.62E−04*
1998	15	0.4810	0.069
1999	26	0.3634	0.068
2007	27	0.1044	0.604
2008	*21*	*0.6661*	*9.80E−04*
2009	*29*	*0.4504*	*0.014*
2010	*45*	*0.3982*	*0.007*
2011	*24*	*0.5445*	*0.006*
2012	*16*	*0.6591*	*0.005*

Note. Years with significant temporal trends in τ_c are italicized.

of 82% and maximum increase 245%). During one year, 1992, τ_c decreases significantly through the summer months (Figure 1a and Table 1).

To determine whether these seasonal trends could result from random variations, we compare the observed time series and 10,000 synthetic time series. We generated each 19-year synthetic time series, with a sample size and event spacing identical to the field data, by randomly sampling τ_c values from a normal distribution with a mean of 192 Pa and a standard deviation of ±39 Pa, which approximates the observational data well. We calculated the Pearson correlation coefficient (R) on each year of this synthetic data series separately, as was carried out on the observational time series, to determine the proportion of years with significant temporal trends. We then perform a Kolmogorov-Smirnov test on each synthetic set of 19 R values against the distribution of R values from the observational data set to determine whether the trends in the field and synthetic data are statistically indistinguishable. Less than 0.001% of 10,000 randomly generated time series reflect the seasonal patterns in τ_c at the Erlenbach (Figure 1b). We conclude that it is highly unlikely that the temporal evolution in τ_c observed across the 19-year record is a result of random variations in τ_c.

Autocorrelation analysis reveals that τ_c remains significantly correlated over 13 transport events from 1987 to 1999 and over 10 transport events from 2007 to 2012 (Figure 1c). This long-lived, interevent correlation illustrates that small, systematic changes in the onset of motion result in significant change in τ_c each year, rather than a single event dominating the signal. In both cases, the number of correlated events is less than the average number of events per year (20 events), indicating that history dependence does not persist across multiple years. Indeed, in the winter months, τ_c appears to be reset to a minimum value, which may be related to riverbed freezing or increased landslide activity (Rickenmann, 1997). However, this study only focuses on the trends observed from February to September.

3.2. Effects of Prior Flow Magnitude on the Onset of Motion

We hypothesize that variation in τ_c is due to reorganization of the channel bed, during interevent flows (Masteller & Finnegan, 2017; Ockelford & Haynes, 2013) and exceptional, transporting flows (Turowski et al., 2009; Yager et al., 2012). We expect τ_c to increase after low shear stresses (Haynes & Pender, 2007), whereas, after high flows, we expect τ_c to decrease (Turowski et al., 2009). To explore this dependence, we focus on the effects of interevent flows without measurable transport and events with measurable bedload separately.

We classify an interevent period as any time separating two transport events absent of any measured coarse (>1 cm) transport and explore the relationship between τ_c and the mean stress of the prior interevent period, $\tau_{\text{i-e}}$. We identify a natural break in the data and find that for interevent flows <104 Pa (corresponding to a discharge $Q = 0.14$ m^3/s), there is no correlation between τ_c and $\tau_{\text{i-e}}$ ($R = -0.019$, $p = 0.764$; Figure 2). For flows exceeding 104 Pa, there is a strong dependence of τ_c on the prior interevent flow magnitude, where a larger $\tau_{\text{i-e}}$ results in an increased τ_c ($R = 0.69$, $p = 1.6 \times 10^{-23}$; Figure 2a). We do not observe any dependence of τ_c on prior interevent duration ($R = 0.54$, $p = 0.270$).

To determine the role of coarse sediment transport in influencing τ_c, we also explored the dependence of τ_c on the peak boundary shear stress of the previous transport event, τ_{peak} (Figure 2b). We choose to explore τ_{peak} to focus on the effects of the most extreme discharges the channel experiences and capture times when the rearrangement or destruction of channel bed structure by sediment transport should be most intense. Due to the variation in τ_c through the record, there is some overlap in the values of $\tau_{\text{i-e}}$, τ_c, and τ_{peak}, such that if they are within the distribution of τ_c, transport does not always occur at shear stresses equivalent to τ_{peak}. However, the maximum values of τ_{peak} far exceed $\tau_{\text{i-e}}$ and τ_c values (Figure S3). Transport events where τ_{peak} exceeds 340 Pa (return period of 1.5 years) do not overlap with the observed distribution of τ_c and

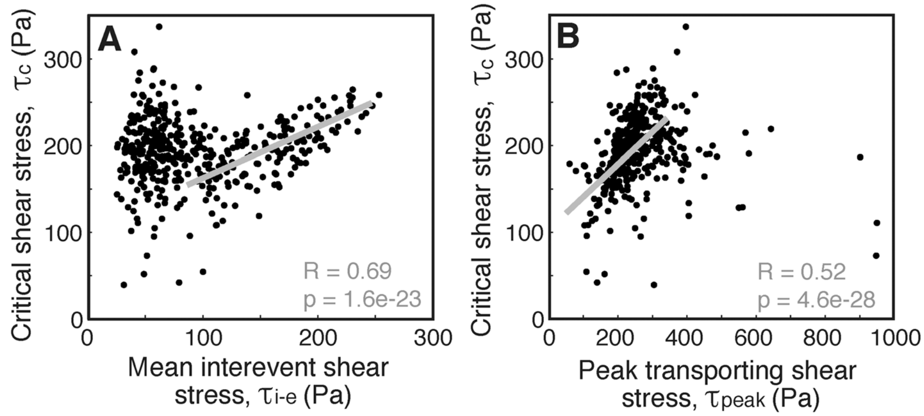

Figure 2. Threshold for motion as a function of (a) mean prior interevent flow magnitude, $\tau_{\text{i-e}}$ and (b) peak transporting flow magnitude of previous transport event, τ_{peak}.

comprise about 15.1% of all transport events. All flows with $\tau_{\text{peak}} > 340$ Pa ($Q = 1.38$ m^3/s), resulted in measurable bedload transport.

When the previous peak transport stress, $\tau_{\text{peak},}$ is below or near τ_c (<340 Pa), similar trends to interevent periods emerge (Figure 2b). In this range, τ_c increases as a function of τ_{peak} ($R = 0.52$, $p = 4.6 \times 10^{-28}$). In contrast, for flows when transport consistently occurs (>340 Pa), there is a significant, inverse relationship between τ_c and τ_{peak} ($R = -0.47$, $p = 0.0025$). However, this signal is dominated by three exceptionally large flows exceeding 900 Pa. Following these three events, τ_c is reduced to a mean value of 124 Pa, or the 5th percentile of critical shear stresses at the Erlenbach. Removing these events from the record, we find a weakly negative relationship between τ_c and τ_{peak} ($R = -0.22$, $p = 0.19$) indicating that any memory of past flows is lost.

4. Potential Mechanisms for Memory Formation and Destruction at the Erlenbach

The strong dependence of τ_c on the magnitude of prior flow across interevent and transporting flows highlights the role of flow history in modifying bedload transport rates. We identify a divergence in the dependence of τ_c on prior flow strength across low-, intermediate-, and high-magnitude flows. The lack of correlation between τ_c and the lowest-magnitude, interevent flows indicates that there is a threshold for the development of history dependence. Below this threshold (~104 Pa), we hypothesize that local rearrangement of the bed cannot occur. Notably, this threshold for the onset of history dependence is distinct from the onset of measurable sediment transport as recorded by the geophone system at the Erlenbach (mean = 192 Pa, std. dev. = 39 Pa).

At the Erlenbach, prior flow magnitude is a main driver of variability in τ_c, whereas prior flow duration appears to have little influence. This lack of correlation between τ_c and interevent duration stands in contrast to previous studies where τ_c is reduced following extended interevent times (Haynes & Pender, 2007; Monteith & Pender, 2005; Reid et al., 1985). In previous work at the Erlenbach, Turowski et al. (2011) compared the discharge at the end of one transport event with the discharge at the start of the following transport event and found that differences were larger when there was an extended period between the two, suggesting a significant role of interevent duration. However, it is likely that the flow magnitude effects that we explore herein also contribute to the previously observed differences in the start and end of transport, though Turowski et al. (2011) did not explicitly explore these. In our analysis, we do observe a secondary effect of interevent duration. For interevent flows shorter than the median interevent flow period (<3 days), there is a strong dependence of τ_c on interevent magnitude ($R = 0.5$, $p = 1.6\text{E}{-14}$). When interevent periods exceed 3 days, these magnitude effects are no longer significant ($R = -0.004$, $p = 0.95$). This loss of magnitude dependence over longer time scales likely occurs because the interevent shear stress varies with interevent duration. Summer discharge variability in the Erlenbach is driven by convective thunderstorms

(Schuerch et al., 2006). Because of this, interevent shear stresses are reduced over longer interevent periods as the channel returns to base flow conditions. Indeed, the average interevent shear stress for interevent periods exceeding three days is 58 Pa—well below the threshold for history dependence that we identify in our analysis.

The onset of the strong dependence of τ_c on the magnitude of intermediate size interevent flows (>104 Pa) reinforces experimental results that gravel bed mobility can be modified in the absence of measurable sediment transport (Masteller & Finnegan, 2017; Ockelford & Haynes, 2013). This positive correlation between $\tau_{\text{i-e}}$ and τ_c is also consistent with previous laboratory experiments, in that higher-magnitude antecedent flows more efficiently stabilize the channel bed (Haynes & Pender, 2007; Monteith & Pender, 2005). Similar dependence is observed for near-threshold, transporting flows, suggesting that gravel beds also reorganize into more stable configurations during low rates of active transport. Indeed, Charru et al. (2004) describes a slow transition from active bedload transport to no observable transport at a constant boundary shear stress as the bed reorganized to a more stable condition. The consistent form of the relationship between τ_c and prior flow strength across interevent and transporting flows suggests that there is a continuum of riverbed behavior (Houssais et al., 2015), rather than a sharp transition as implied by existing bedload transport models.

The loss of history dependence in τ_c following high magnitude flows (>340 Pa) indicates (Figure 2b) that the mechanisms for bed strengthening (e.g., Charru et al., 2004; Masteller & Finnegan, 2017) become less effective as flow magnitude increases well above τ_c. We hypothesize that as the degree of transport increases grains are more effectively removed from stable pockets, resulting in increased disorder, ultimately leading to loss of memory. The dislodging of more stable particles can drive bed dilation (Allan & Frostick, 1999; Houssais et al., 2015; van der Elst et al., 2012), disrupt particle interlocking (Oldmeadow & Church, 2006), and increase the protrusion and exposure of particles to high fluid drag, reducing τ_c (Kirchner et al., 1990; Masteller & Finnegan, 2017). In the Erlenbach, larger-scale channel changes are observed when shear stresses well exceed 340 Pa, including observations of step migration (~580 Pa, Golly et al., 2017) and boulder mobilization (>900 Pa, Turowski et al., 2009). This suggests that while these processes likely add to the disorder on the channel bed, the loss of history dependence in τ_c is more likely related to the onset of reliable bedload transport, which occurs near $\tau_{\text{peak}} = 340$ Pa. Indeed, every year of the transport record lacking a significant seasonal trend experiences τ_{peak} exceeding 340 Pa.

A number of other physical mechanisms may also result in this systematic variation in τ_c, including (I) changes in surface grain size (Chen & Stone, 2008; Moog & Whiting, 1998) or (II) changes in sediment supply (Dietrich et al., 1989; Yager et al., 2012). Gravel pavement development via armoring, or an increase of bed surface grain size, is thought to require significant transport to coarsen the bed surface (Parker & Klingeman, 1982). This potential mechanism cannot be evaluated directly, as there are not time series measurements of transported and bed surface grain size during our study period. However, because τ_c exhibits a dependence on flows without any detectable bedload transport it unlikely that armoring is the dominant process by which τ_c evolves at the Erlenbach during below- and near-threshold flows. However, infrequent, high-magnitude floods, can result in a transient fining, which may contribute, in part, to the observed decrease in τ_c following these extreme events. Indeed, Yager et al. (2012) noted a slight fining following an extreme event in the Erlenbach in 2007. However, bedload flux predictions did not change significantly even when these variations in grain size were accounted for (Yager et al., 2012). Thus, at this site, it may be that changes to the bed surface grain size are too small to affect bedload transport and τ_c.

Additionally, because surface grain size is tightly linked to sediment supply (Buffington & Montgomery, 1999; Dietrich et al., 1989), both bed armoring and sediment supply require significant transport to modify bed stability. Given this, it is also unlikely that variations in sediment supply are related to variations in τ_c following interevent flows. That said, active landsliding triggered by high-magnitude flows (Golly et al., 2017) may reduce τ_c during large floods by delivering relatively disordered particles or fine sediments to the bed (Wilcock, 1998; Wilcock & Kenworthy, 2002). However, the timing of individual landslide events at the Erlenbach is not well documented, making it difficult to ascertain these effects directly.

Previous studies at the Erlenbach reinforce the idea that there is a transition in channel behavior corresponding to the loss of history dependence. Yager et al. (2012) tracked tracer particle activity for various grain sizes during transport events with peak stresses ranging from 187 to 304 Pa. For flows ranging from 187 to 271 Pa, 1% of tracer particles with a diameter of 15.4 cm are transported. During the largest event in the study (τ_{peak} = 304 Pa), the activity of these tracers increases by an order of magnitude to 18%. Tracer particles more comparable to the median mobile grain size at the Erlenbach (D_{tracer} = 7.6 cm, D_{50} = 7 cm from Wyss et al., 2016) exhibit a similar pattern. The active fraction of D_{50}-size tracers increases threefold, from 15% to about 50% during the largest event. The onset of pervasive transport of these particles occurs near the loss of correlation between τ_c and prior flow magnitude observed by this study (~340 Pa), reinforcing the notion that intense sediment transport disrupts memory in τ_c.

Our observations have parallels with granular flows under shear, namely, the transition from strengthening to weakening behavior with increasing applied shear stress. Low shear stresses result in compaction and an apparent increase in effective friction (Boyer et al., 2011; van der Elst et al., 2012) and high shear stresses result in dilation and a decrease in friction (Jaeger et al., 1996; van der Elst et al., 2012). This magnitude-dependent response of granular media has also recently been observed experimentally under laminar flow by Houssais et al. (2015). Allan and Frostick (1999) also noted dilation of the coarse portion of a mixed-size, experimental bed at the onset of bedload transport. Observations from the Erlenbach exhibit similar trends to those observed in idealized experimental conditions, suggesting that the same transition between compaction and dilation behavior may be relevant for natural channels composed of nonspherical grains exposed to turbulent flow.

This similarity across experimental and natural systems speaks to the importance of flow history effects, even in complex natural channels like the Erlenbach, with a wide grain size distribution, significant hillslope-channel coupling, and persistent structures like channel steps. This strong history dependence, consistent with observations from low gradient systems (e.g., Haynes & Pender, 2007; Reid et al., 1985), suggests some universality in the processes that drive the observed variability in the threshold for sediment motion. In addition, even in the Erlenbach, a complex, high gradient, step-pool channel, we find that the shear stress thresholds for the formation and loss of memory in τ_c appear to be most closely related to the intensity of coarse particle transport. This implies that history dependence in τ_c is likely to arise in gravel channels of many forms. We expect that similar patterns of stabilization under low and intermediate flows and disruption of that stabilization under high flows should persist, though the details of this dependence may vary (Rickenmann, 2018). For example, previous results from Turowski et al. (2011) indicate that interevent flows are less effective in modifying channel stability in proglacial streams. However, more work is required to explore the sensitivity of gravel beds to these magnitude effects as a function of grain size distribution, slope, and other channel characteristics.

5. Conclusions

Using a long-term record of bedload transport from a steep mountain stream, we observe a strong dependence of τ_c on the magnitude of previous flows, when the previous flow shear stress is between 104 and 340 Pa. This dependence results in a pattern of seasonally increasing τ_c. Channel stabilization occurs across times with weak and medium bedload transport and times without transport, suggesting that local rearrangement of grains build memory in τ_c. This history dependence is disrupted following events with peak stresses exceeding 340 Pa, where bedload transport reliably occurs. We infer that this loss of memory is associated with significant mobilization of coarse particles, reorganizing the bed. This finding is consistent with previous work on stress history and has parallels to the evolution of sheared granular media, where particle interlocking, material strength, and memory are linked.

Similar to previous studies, our findings at the Erlenbach strongly suggest that the assumption of a constant τ_c may not accurately represent sediment transport dynamics, as stress history effects may be universal across many types of fluvial systems. Our results provide support for the treatment of the onset of motion as a state-dependent quantity (Johnson, 2016). We identify both a range of flows over which there is a strong dependence on prior history and the flows that cause this dependence to break down. These results suggest that the development and implementation of a state function for τ_c, which incorporates flow history effects and hydrograph variability, has potential to improve predictions of sediment mobility.

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