Key Points:

- Despite constant flow and little change of surface grain size, critical shear stress changed in response to bed surface state
- Bed surface structures increased the shear stress required to entrain sediment
- The Shields Number is a bed state parameter

Supporting Information:

- Supporting Information S1

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Experimental Insights Into the Threshold of Motion in Alluvial Channels: Sediment Supply and Streambed State

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Abstract Flume experiments were conducted to understand the role of episodic sediment supply on the threshold of motion, streambed state, and stability in gravel bed channels. We demonstrate the importance of bed surface evolution on grain entrainment and hence channel stability. The grain-size distribution (GSD) of the bed surface coarsened quickly under no feed conditions to develop an armored surface. Armor continued to experience particle exchange but remained intact throughout the experiment, and was relatively insensitive to subsequent sediment supply pulses. Bed surface structures (coarse-grain clusters, stone cells, and imbrication) developed during the experiments and continued to respond dynamically to sediment pulses. Thresholds of motion estimated by various methods increased during the experiment and fluctuated in response to changes in sediment supply, though they were poorly predicted either by surface grain-size changes or by surface structure. This reinforces the idea that the threshold for motion is not constant in poorly sorted gravel-bed streams, and it confirms that the critical Shields number in transport models represents a bed state parameter, not a grain mobility number. Changes are controlled by a variety of factors including sediment supply regime, the degree of bed structuring, and the history of bed evolution. Our results highlight the importance of sediment supply regime as a control on bed surface evolution and the channel stabilizing function played by surface structures.

1. Introduction

Multiple bed surface characteristics, including surface grain size distribution (GSD), grain protrusion and surface roughness, and development of coarse-grain clusters, respond to water and sediment supply in ways that generally enhance bed stability. As the sediment supply decreases, the gravel-bed surfaces coarsen and bed stability increases (Dietrich et al., 1989; Gessler, 1970; Nelson et al., 2009). This armoring (surface coarsening relative to the subsurface or the sediment supply) occurs when transport capacity exceeds sediment supply, resulting in an immobile armored bed surface. Armoring can also reflect a bed adjustment so that the GSD of sediment transported out of a reach matches the sediment supply GSD from upstream, producing a mobile armor (e.g., Parker & Klingeman, 1982; Temple & Wilcock, 2005).

Early work on the organization of gravel beds showed that the process of bed coarsening is accompanied by the formation of structural features, including clustering of the relatively large and less mobile grains that further reduces the overall mobility of the bed surface sediments and increases roughness (e.g., Brayshaw, 1984; Brayshaw et al., 1983). A wide variety of bed structures is now known to occur that form coherent patterns of clasts in gravel-bed rivers (see review in Venditti et al., 2017). These features are larger than individual clasts and generally smaller than channel-scale features (e.g., bars or step-pool features). The bed features include clusters (e.g., Brayshaw, 1984; Hassan et al., 2020; Strom & Papanicolaou, 2009), stone lines (Curran & Tan, 2014; Laronne & Carson, 1976) or transverse ribs (e.g., Allen, 1984; Koster, 1978), and reticulate stone cells (e.g., Church, et al., 1998; Hassan & Church, 2000). The most easily recognized and identified structures are clusters, which are closely nested groups of clasts aligned roughly parallel with the flow and often formed by a large keystone around which smaller particles collect (Venditti et al., 2017). They tend to develop rapidly in gravel-bed streams and may remain largely intact even during relatively large flow events (Lamarre & Roy, 2008). The impact of sediment supply on bed surface structuring has been explored experimentally (e.g., Hassan & Church, 2000; Johnson, 2017; Qin et al., 2017; Vazquez-Tarrio

et al., 2020). Like armor layers, the development of gravel-bed structures is affected by sediment feed rates relative to transport capacity, and influences the amount of shear stress available to transport sediment. For example, Hassan and Church (2000) showed that structures are formed as a sequence (clusters-stone lines-stone cells) with decreasing sediment supply and that between 17% and 47% of bed shear stress was carried by bed surface structures. Coarse grains on the surface are a key control on sediment mobility, hence on channel morphology and stability (Church & Ferguson, 2015; Harrison, 1950; Mackenzie & Eaton, 2017; Parker, 1978). A relatively small number of studies have looked at interactions among coarsening, surface topography, and sedimentary structures, and there is evidence that the temporal evolution and relative importance of these factors in controlling bed stability vary with initial conditions and the history of flow and transport conditions (e.g., Johnson et al., 2015; Johnson, 2017; Masteller & Finnegan, 2017; Perret et al., 2018, 2020).

The threshold of gravel motion has long been recognized to be influenced by parameters other than the mean shear stress and the median grain size, as originally formulated in the classic work of Shields (1936). Field and laboratory studies reveal wide scatter in the relation between sediment transport rate and flow characteristics, which results in wide scatter in the threshold of motion (Buffington & Montgomery, 1997). This variability has been attributed to many interrelated factors, including bed surface coarsening (e.g., Andrews & Parker, 1987; Gessler, 1970; Parker et al., 1982; Parker & Klingeman, 1982; Tait & Willetts, 1992), sand or finer gravel on the bed (Perret et al., 2018; Venditti et al., 2010a, 2010b; Wilcock & Crowe, 2003), protrusion of grains into the flow relative to the rest of the bed (e.g., Fenton & Abbott, 1977; Hodge et al., 2020; Kirchner et al., 1990; Masteller & Finnegan, 2017; Perret et al., 2020; Roberts et al., 2020), intergranular friction and imbrication (Johnston, 1922; Yager et al., 2018), the geometry of pockets from which grains are entrained (Buffington et al., 1992; Johnston & Andrews, 1998; Kirchner et al., 1990; Masteller & Finnegan, 2017), influence of bed surface structures (e.g., Brayshaw, 1984, 1985; Brayshaw et al., 1983; Church et al., 1998; Cin, 1968; Laronne & Carson, 1976), bedform evolution (e.g., Recking et al., 2009; Saletti et al., 2015), and sediment availability (e.g., Andrews & Parker, 1987; Hassan & Church, 2000; Mao et al., 2011; Parker et al., 1982; Recking, 2012). These factors cause spatial and temporal changes in the local threshold for motion (e.g., Hodge et al., 2020; Johnson, 2016; Masteller et al., 2019; Turowski et al., 2011; Yager et al., 2018). It has been shown that the threshold of motion increases as grain protrusion decreases (Masteller & Finnegan, 2017), although they have argued that pocket geometry is controlling thresholds for individual grains (Hodge et al., 2020). Given this very large number of controls, the gravel threshold of motion can be thought of as a lumped parameter that is controlled by the “bed state,” a somewhat loosely defined term to describe the ensemble of factors that control bed mobility and transport rates. Johnson (2016) argued that the threshold of motion parameter in bedload transport models should be conceptualized as a nonunique state variable for channels that characterizes not only initial grain mobilization but also the ease of transporting sediment for flow intensities exceeding threshold conditions.

The formation of gravel-bed structures has been shown to influence the Shields number required to entrain gravel particles. In a combined field and experimental study, Church et al. (1998) showed that the joint effect of armoring (grain coarsening) and bed surface structures increased the critical Shields number up to 0.09, about twice the value of ~ 0.045 commonly used for poorly sorted gravel (e.g., Komar & Li, 1988; Miller et al., 1977; Wilcock & Southard, 1989). Oldmeadow and Church (2006) further reported that the critical Shields number for a reach in a small gravel-bed stream, where they had manually destroyed the surface structures, was reduced by 18% relative to a reference reach in which all structures remained intact. Similarly, Hendrick et al. (2010) attributed larger Shields stress values (0.08) to stone clusters and reported lower values (0.06) associated with uniform bed material with no clusters. In contrast, Piedra et al. (2012) found that, while distributed large particles indeed increased bed stability, clustered large grains were less effective at stabilizing the bed surface.

The critical shear stress required to initiate sediment motion can be characterized in different ways. One way is through direct observations of the flow conditions that cause grains to move, but this threshold can be difficult to define because it requires subjective definition of what constitutes grain motion of a gravel bed (one grain wiggles, one grain moves, and general grain motion). Others have used empirical fits to observe the flow, shear stress, or stream power at which sediment transport becomes negligibly small (e.g., the reference shear stress [Parker et al., 1982]). More physically based models have been developed

Figure 1. Experiment setup showing the locations of the water supply, sediment feed, fixed bed, mobile bed, observation reach, and light table.

in which thresholds of motion are expressed in terms of grain protrusion and intergranular friction (Yager et al., 2018), or the threshold is considered as a distribution of values (Paintal, 1971; Wong & Parker, 2006). Others have modified the reference shear stress threshold parameter in the Wilcock and Crowe (2003) transport model to evolve through time as a function of sediment supply, erosion, and deposition (e.g., Johnson, 2016). Recently, Perret et al. (2018, 2020) conducted novel experiments exploring bed stability as a function of surface roughness and grain packing, and found a 5%–19% reduction in critical threshold stress (a reference stress) could be attributed to increased grain packing from different initial conditions. Interestingly, they found a relatively weak dependence of their reference stress on bed surface roughness statistics, although surface statistics did more strongly correlate with transport statistics through a shear stress exponent. Thus, while previous work suggests that effects of bed structure, sediment supply, and other influences on bed stability should influence thresholds of motion, isolating and quantifying these controls remains elusive.

The overall goal of the present work is to better understand how bed surface grain size, structure development, and the threshold of motion simultaneously adjust to sediment supply perturbations in a gravel-bed river. A secondary goal is to quantitatively understand how bedload transport rates are influenced by these stabilizing bed adjustments. To explore how thresholds of motion evolve in conjunction with measurable bed surface characteristics, we use recent experiments that examined the impact of sediment pulses on sediment transport rate, sediment storage, the evolution of surface structures, and bed roughness statistics (Elgueta-Astaburuaga et al., 2018; Elgueta-Astaburuaga & Hassan, 2017, 2019; Hassan et al., 2020; Müller & Hassan, 2018). We explore the following hypotheses at channel – as opposed to granular – scale: (1) thresholds of bed material motion change systematically as surface armoring and coarse-grain structures develop; (2) bed material transport is regulated in the partial transport regime by bed structuring; and (3) the history of previous sediment supply perturbations affects threshold and channel response.

Figure 2. Grain-size distribution of the bulk sediment mixture used in the experiments.

2. Experimental Design and Methods

The experiments were carried out at the Mountain Channel Hydraulic Experimental Laboratory at the University of British Columbia. The flume and the experiments have been described in Elgueta-Astaburuaga and Hassan (2017, 2019), Elgueta-Astaburuaga et al. (2018), Hassan et al. (2020), and Müller and Hassan (2018). The experiments were conducted in an 18-m-long, 1-m-wide, and 1-m-deep tilting flume (Figure 1). The working flume length was 12 m with an initial bed slope of 0.0218 m/m. The flume was filled with a 10-cm layer of poorly sorted sediment ranging from 0.5 to 64 mm (Figure 2). For visualization and particle size identification, each half-phi grain-size class was painted in a distinct color.

The water discharge was kept constant at 0.065 m³/s, sufficient to move all sediment sizes at least partially (*sensu* Wilcock & McArde, 1993),

Figure 3. (a) Sediment transport rate and feed rate versus time. Red markers show the time at which different measurements were made: (Photo/WSE) bed surface photos, water surface elevation and bed elevation, and (Scan) bed elevation laser scans. Sediment transport rates were measured every second and they are plotted as gray points. Black line is a 10-minute average. Gaps of missing data in R1 and R2 are due to technical problems with the light table. (b) Cumulative mass transport through time for different grain-size classes. For simplicity, the missing data periods during R1 and R2 are plotted as times with zero transport.

and for the bed to armor within the first experimental run. Seven runs, each lasting 40 h, were conducted with different sediment supply regimes (Figure 3). Runs R1 and R7 were conducted with no sediment feed. Runs R2–R6 each had the same cumulative mass of sediment supply (300 kg total per run), but over different time intervals. Runs R2 and R6 had a constant feed (2.0 g/s/m), R3 had one pulse of sediment feed (83 g/s/m during the first 40 min), R4 had four pulses (83 g/s/m during the first 15 min every 15 h), and R5 had two pulses (83 g/s/m during the first 20 min every 20 h) (Table 1). Sediment identical in GSD to the initial bulk mixture was fed into the flume at the upstream end of the mobile reach using a conveyor belt (Figure 1). Transported material passed over a light table and was trapped in a 0.18-mm mesh screen at the end of the flume. By holding flow constant in a straight channel with fixed walls and by maintaining a constant sediment input in each run, we reduced the variables governing sediment transport rates to temporal variations in sediment supply (the upstream boundary condition) and bed surface evolution (autogenic system dynamics).

Water depth and bed and water surface slopes were measured along the 12-m working section at 0.5-m intervals with a variable temporal resolution as shown in Figure 3; measurements were taken more frequently after each pulse to detect changes caused by episodic sediment supply. High-resolution images of sediment crossing a light table were used to determine sediment flux and particle size of material exiting the flume (Elgueta-Astaburuaga & Hassan, 2017; Zimmermann et al., 2008). Data to measure bed surface GSD and topography were made periodically in a 2-m-long observation reach between 6 and 8 m from the flume outlet (Figure 1) after temporarily draining the flume. Pictures of the bed were used to document grain-size changes and map surface structures, and bed elevation was measured at 2-mm spacing using a green laser scanner to determine surface roughness, bed surface structures, and sediment storage. The bed surface GSD

Table 1
Summary of Hydraulics and Sediment Data

Run	Sediment feed regime	Magnitude of each pulse (kg)	Pulse recurrence interval (h)	Feed rate (g/s/m)	Water depth (cm)	Water surface slope (m/m)	Bed slope (m/m)	Mean bedload transport rate for the run (g/s/m)	Standard deviation of transport rate	D_{50s} (bed surface)	D_{50l} (bedload)
R1	No feed	-	-	0	-	-	0.017	1.29	27.06	17.0	4.1
R2	Constant	-	-	2	7.3	0.017	0.016	0.65	5.12	17.5	4.7
R3	1 pulse	300	40	83 ^a	8.0	0.016	0.018	1.56	7.64	16.8	2.3
R4	4 pulses	75	10	83 ^b	8.3	0.018	0.020	0.98	3.17	16.1	4.1
R5	2 pulses	150	20	83 ^c	7.2	0.020	0.022	1.19	10.32	16.1	2.8
R6	Constant	-	-	2	7.5	0.019	0.022	1.25	31.32	16.1	3.2
R7	No feed	-	-	0	7.3	0.020	0.022	0.42	1.56	17.4	2.8

Note. For more details, see Figure 3.

^aSediment was fed within the first hour (300 kg/1 h). ^bSediment was fed within the first 15 min of each (75 kg/15 min). ^cFed within the first half an hour (150 kg/30 min).

in the observation reach was estimated by overlaying a 5-cm grid over each photograph and identifying particle size at the grid intersections. Individual grains were identified by color for fractions coarser than 2.83 mm; material finer than 2.83 mm was difficult to distinguish and therefore was lumped into one size class. To quantify roughness, the standard deviation of the bed elevation was calculated for each 2-m observation reach digital elevation model (DEM). We emphasize that detailed measurements of clustering and grain size were made only in the observation reach, not made in the downstream 6 m of the flume.

3. Data Analyses

3.1. Change in the Surface Coverage

We estimated how the proportions of coarser size fractions changed on the bed surface through time (a control on particle/bed mobility) in response to changes in sediment supply. To do so, we used image analysis to automatically measure areas occupied by the larger grains by color. This method captured and was accurate for particles sizes >11 mm (Figure S1). This corresponded to the five coarsest grain-size classes in the sediment mixture, representing all sizes larger than the D_{65} of the bulk GSD (Figure 2) and the D_{826-40} of the surface GSD (Elgueta-Astaburuaga & Hassan, 2017). Images covering the observation reach were used to evaluate variation in bed surface GSD. Additional details are provided in the Supporting Information (Supporting Information S1; Figures S1 and S2).

3.2. Identifying Particle Structures

We identified discrete grain structures using a semi-automated multifactor method described in Hassan et al. (2020). Here we provide a brief summary. First, using the color code, all potential anchor particles (> D_{80} of the bulk distribution – corresponding to 20 mm) exposed on the bed surface were identified. Second, the relative exposure (elevation of the potential particle above the average local bed elevation; the “projection” of Kirchner et al., 1990) was calculated using the DEMs. Third, a structure was identified if the potential anchor particle had at least three clasts larger than 8 mm (identified by color code) in contact with it. Finally, the boundary of each structure was delineated. All particles larger than 8 mm in contact with each other and the anchor particles were encompassed in a structure.

3.3. Estimating Thresholds of Motion

3.3.1. Thresholds of Motion from Transport Rates

We used two existing bedload transport models to evaluate how thresholds of motion evolve through time in these experiments, based on shear stress, grain size, and transport rate data. These threshold parameters are defined by their respective models. First, we back-calculated the nondimensional threshold for motion, τ_c^* , in the Wong and Parker (2006) model (hereafter referred to as W&P), which is based on the original Meyer-Peter & Müller (1948) function and data. We used Equation 24 in W&P that maintains the exponent equal to 1.5:

$$q_b^* = 3.97(\tau^* - \tau_c^*)^{1.5}, \quad (1)$$

wherein W&P proposed $\tau_c^* = 0.0495$. Dimensionless sediment transport rate q_b^* (Einstein number) and dimensionless shear stress τ^* were calculated from experimental data using standard relations:

$$q_b^* = \frac{q_s}{D \sqrt{\frac{\rho_s - \rho}{\rho} g D}}, \quad (2)$$

$$\tau^* = \frac{\tau}{(\rho_s - \rho) g D}, \quad (3)$$

where q_s is dimensional volumetric bedload transport rate per unit width, D is grain diameter, g is gravitational acceleration (9.81 m/s^2), and ρ_s ($2,650 \text{ kg/m}^3$) and ρ ($1,000 \text{ kg/m}^3$) are sediment and water densities, respectively. We evaluated the bed shear stress τ using the depth-slope product (i.e., $\tau = \rho g d S$, where d is the water depth and S the water surface slope). For grain size (D), we used the median size of the transported material (D_{50}).

Second, following the method described by Johnson (2016), we also back-calculated the threshold of motion parameter from the Wilcock and Crowe (2003) surface-based transport model (hereafter referred to as W&C), which calculates separate transport rates for different grain-size classes. The W&C model uses a “reference bed stress” that describes a very low transport rate, rather than an actual threshold for entrainment (Parker, 1990). W&C defined a dimensional reference stress τ_{ri} for every size class i , with dimensionless equivalent $\tau_{ri}^* = \tau_{ri} / (\rho_s - \rho) g D_i$, where D_i is the mean diameter of size class i . The W&C threshold parameter τ_{rs50}^* is the reference Shields stress for the median surface grain size (D_{s50}). The original W&C model assumes that τ_{rs50}^* varies between 0.021 and 0.036 as a specific function of sand content. Instead, we empirically solve for τ_{rs50}^* using experimental observations of transport rates for each size class, rather than imposing the sand content dependence of W&C.

The W&C hiding function (modified slightly as described in Johnson [2016]) determines how thresholds for different grain sizes vary relative to the median surface size:

$$\frac{\tau_{ri}^*}{\tau_{rs50}^*} = \left(\frac{D_i}{D_{s50}} \right)^{b-1}, \quad (4)$$

$$b = \frac{0.67}{1 + e^{\left(1.5 - \frac{D_i}{D_{s50}}\right)}}. \quad (5)$$

The W&C model applies these threshold parameters to relate nondimensional transport rates for each size class to driving shear stresses using

$$W_i^* = \begin{cases} 0.002 \left(\frac{\tau}{\tau_{ri}} \right)^{7.5} & \text{for } \frac{\tau}{\tau_{ri}} < 1.35 \\ 14 \left[1 - \frac{0.894}{\left(\frac{\tau}{\tau_{ri}} \right)^{0.5}} \right]^{4.5} & \text{for } \frac{\tau}{\tau_{ri}} \geq 1.35 \end{cases} \quad (6)$$

where W_i^* is a nondimensional bedload transport rate for grain-size class i , defined as

$$W_i^* = \frac{(\rho_s / \rho - 1) g q_{bi}}{F_i u_*^3}, \quad (7)$$

where q_{bi} is the volumetric transport rate per unit channel width of grains of size i , F_i is the fraction of size i on the bed surface, and $u_* = \sqrt{\tau / \rho}$ is the shear velocity.

We used regressions to empirically calculate τ_{rs50} in two ways. Both methods use our experimental measurements of q_{bi} , F_i , and u_* to calculate W_i^* for each time step (Equation 7), and then apply Equation 6 to calculate τ_{ri} for each size class. The first method assumes that the W&C hiding function exponent is correct and uses Equation 5 to calculate b . The second approach uses nonlinear regressions (the built-in Matlab “fit” function for “power1” power laws, using default values for other Matlab regression parameters) to simultaneously calculate best fit values for b and τ_{rs50}^* in Equation 4, and does not use Equation 5. Both methods provide regression-based confidence intervals on τ_{rs50}^* .

3.3.2. Threshold of Motion From Grain-Size Ratio

We also adopted an approach proposed by Church et al. (1998) and Hassan and Church (2000) to infer thresholds of motion from surface coarsening. Dietrich et al. (1989) defined a dimensionless sediment transport rate parameter q_* to relate surface coarsening and bedload transport rates:

$$q_* = \left(\frac{\tau_b - \tau_{cs}}{\tau_b - \tau_{cl}} \right)^{1.5} \quad (8)$$

where τ_b is the mean bed shear stress calculated using the depth-slope product, and τ_{cs} and τ_{cl} are critical shear stresses calculated for the median bed surface grain diameter (D_{50s}) and the median grain diameter of the bedload (D_{50l}), respectively.

The idea underlying Equation 8 is that, as q_* approaches 1, the GSD of the bed surface is the same as that of the subsurface (i.e., no armoring or coarsening) because $\tau_{cl} \sim \tau_{cs}$. Equation 8 may alternatively be interpreted as a metric of the transport rate when the bed is not coarsened, compared to when the bed has armored. As the bed becomes increasingly armored, τ_{cs} approaches τ_b and $q_* \rightarrow 0$. In fact, Church et al. (1998) measured in their set of no-feed armoring experiments a ratio of $q_* \approx 0.001$. Under these conditions of $q_* \approx 0$, Equation 8 can be simplified to the following relation (see Equation 2 in Dietrich et al. [(1989)]):

$$\frac{\tau_b}{\tau_{cl}} \approx \frac{D_{50s}}{D_{50l}}. \quad (9)$$

Thus, when the channel bed is fully armored, that is, when $q_* \rightarrow 0$, Equation 9 can be used to back-calculate the value of τ_{cl} if the bed shear stress and ratio D_{50s}/D_{50l} are known. Alternatively, the grain-size ratio D_{50s} / D_{50l} can be predicted by dividing the mean bed shear stress by the critical shear stress expected for the bulk sediment D_{50} (assuming that the transported load approximates the GSD of the supply). Bed structuring has not been accounted for in developing Equations 8 and 9 from standard bedload transport models (Dietrich et al., 1989). Therefore, comparing the predicted and experimentally measured D_{50s} / D_{50l} ratios provides an indirect measure of the degree of bed surface structuring.

We used the sediment transport rate measured near the beginning of R1 (after 1 h of run time) as the condition associated with negligible surface coarsening (see Church et al., 1998). In that case, the sediment transport rate can be expressed as $q_b \propto (\tau_b - \tau_{cl})^{1.5}$, where τ_{cl} can be taken as the nominal critical Shields stress for a gravel bed. To be consistent with the assumptions and analyses of Dietrich et al. (1989), Church et al. (1998), and Hassan and Church (2000), we assumed that the critical Shields stress for loose sediment with negligible structures is equal to $\tau_{cr}^* = 0.045$ (see also Komar & Li, 1988; Miller et al., 1977; Wilcock & Southard, 1989).

4. Results

We first show how transport rates responded to sediment feed pulses and calculate the nondimensional threshold for motion, τ_c^* , that correspond to these transport rates. To better understand how the interplay of multiple variables may control the threshold of motion, we then present corresponding data on how the bed surface evolved through time, including the creation and destruction of coarse-grain clusters.

4.1. Sediment Transport

The experiment started with 40 h without sediment feed (R1). Sediment transport rates were initially high and then rapidly declined by roughly an order of magnitude as the initially placed bed stabilized (Figure 3). R2–R6 each had the same cumulative mass of sediment fed into the flume, but over different time periods within each run. Feed rate varied between the constant rate and pulsed feed episodes (Table 1). Transport rates increased gradually, then tended to level off during constant feed runs R2 and R6. In R3, the introduction of a large, 300-kg pulse of sediment at 80 h resulted in a sharp increase in the sediment transport rate (measured at the downstream end) in less than 1 h, after which the transport rate declined rapidly. Similar trends but with smaller peak transport rates were obtained for R4 (four 75-kg pulses) and R5 (two 150-kg pulses). When the feed was stopped (R7), the transport rate gradually decreased and then roughly stabilized for the duration of the run, but with a much smaller decrease than in no-feed R1. Additional analysis of sediment transport for this experiment has been published (Elgueta-Astaburuaga et al., 2018; Elgueta-Astaburuaga & Hassan, 2017; Müller & Hassan, 2018).

Grains smaller than 8 mm ($\approx D_{60}$ of the bulk sediment; $\approx D_{16}$ of the surface after armor development) remained quite mobile throughout the experiment, while coarser size classes exhibited partial mobility (Elgueta-Astaburuaga et al., 2018). All size classes had a significant increase in transport rates measured at the downstream outlet in response to the largest sediment pulse (R3) (Figure 3b). R4 and R5 sediment pulses also caused measurable increases in transport rates of all but the largest size classes shown (> 32 mm). Fractions > 32 mm constituted about 5% of the bulk (feed) sediment; during the 280-h experiment, 95% of these grains remained upstream from the observation area.

4.2. Thresholds for Motion

Figure 4 shows time series of entrainment thresholds back-calculated from the W&P model (Equation 1), from the W&C transport model (Equations 4–7), and from the Church et al. (1998) model (Equation 8 and 9). The W&P model calculation uses the D_{50l} of the transported sediment for nondimensionalizing stresses and thresholds, and does not incorporate surface grain-size changes. The W&C and Church et al. (1998) models use both transported and surface GSDs in their determination of thresholds. Figure 4b uses the W&C hiding function to calculate exponent b (Equation 5), while Figure 4c instead uses nonlinear regression to simultaneously find best fit values for τ_{rs50}^* and exponent b without imposing Equation 5. Both the overall similarity in τ_{rs50}^* estimates (Figures 4b and 4c) and the consistency between Equation 5 and the regression-based estimates of exponent b (Figure S3) suggest that the W&C hiding function is broadly consistent with these transport data. Importantly, the similar trends in Figures 4a–4d suggest that our results are not strongly dependent on the choice of bedload transport model. The linear correlation between the W&P and W&C thresholds has $R^2 = 0.69$ (Figure S4). The two W&C methods are correlated with $R^2 = 0.81$.

Figure 4. Nondimensional thresholds of motion calculated with different methods. (a) Time series of τ^*_c calculated following Wong and Parker (2006) and Lamb et al. (2008; $\tau^*_c = 0.15 S^{0.25}$, where S is the bed surface slope). (b) Time series of reference shear stress using the Wilcock and Crowe (2003) transport model (Equations 4, 6, and 7), including their hiding function (Equation 5). (c) Using Equations 4, 6, and 7 but not Equation 5. Error bars in (b) and (c) are 95% confidence intervals. (d) Time series of τ^*_c calculated following Church et al. (1998) and Hassan and Church (2000), using Equations 8 and 9, with the reference critical value of 0.045. (e) Time series of the bed shear stress calculated using the slope-depth product.

Figure 5. Einstein number q_b^* versus Shields number τ^* for measured transport rates (points) and Wong and Parker (2006) Equation 1 using different values for τ_c^* (dashed lines). For comparison, the dimensionless transport rate of $q_b^* = 0.002$ (“threshold” reference value) of Parker et al. (1982) is plotted.

The overall experiment started with a low τ_c^* , around 0.04 (W&P) after the initial bed erosion of no-feed R1. During the constant-feed R2, thresholds either increased modestly (to around 0.045 for both W&P and Church et al. (1998), τ_c^*) or stayed relatively constant (for W&C, τ_{rs50}^*). The single large sediment supply pulse at the start of R3 caused an overall increase in threshold, apparently rapid at first and then leveling off.

Differences in threshold response calculated using the different methods are observed right after the sediment pulses. For the W&P model, the increase is observed in the very first measurement taken 1 h after the pulse (i.e., the change happens between 80 and 81 h). However, a difference that appears subtle in Figure 4 is that the W&C methods do not show a threshold increase until 4 h after the pulse (i.e., 81 and 82 h still have low thresholds, and the increase happened between 82 and 84 h). Later and smaller sediment pulses of R4 and R5 tend to cause temporary threshold increases in the W&P model (Figure 4a). In contrast, in the W&C model, the R4 and R5 sediment pulses tend to cause temporary threshold decreases, particularly evident in Figure 4c. The temporal evolution of the threshold of motion back-calculated with the approach proposed by Church et al. (1998) and Hassan and Church (2000) (Equations 8 and 9, shown in Figure 4d) is very similar to the one obtained using W&P, except

for a peak around 80 h, just after the large sediment pulse in R3. Thus, the different threshold models do predict somewhat different immediate responses to the sediment pulses.

A significant feature of the threshold stresses is that they correlate strongly with the total bed stress (Figure 4e). We find the correlation with τ_c^* (as in W&P, Equation 1) has $R^2 = 0.91$. For τ_{rs50}^* (as in W-C calculated by regression, Equation 4), $R^2 = 0.65$, while for τ_c^* (as in Hassan & Church, 2000, Equation 9), $R^2 = 0.85$. Hence the threshold tracks total stress. The lower correlation for τ_{rs50}^* presumably reflects the different computed initial response to the sediment pulses.

Figure 5 shows that the measured sediment transport rates were relatively low; in most cases, they were below a dimensionless transport rate of $q_b^* = 0.002$, a value frequently adopted as a “threshold” reference value (Parker et al., 1982). The dashed lines are contours of τ_c^* calculated using the W&P model. Only two measurements associated with larger sediment pulses (R3 and R5) plot above $q_b^* = 0.001$. Most transport in these experiments occurred at stresses only slightly above the threshold values, indicating that even large pulses of sediment resulted in relatively rapid transitions back to low transport rates.

4.3. Bed Surface Slope, Grain Size, and Structure

We next explore the evolution of reach slope, surface grain size, and bed structure to better understand how thresholds of motion may correspond with measurable bed changes (Figure 6). At the beginning of R1 the bed surface abruptly degraded, decreased in slope (Figure 6b), and coarsened (Figure 6c). During R2 with constant sediment feed, the bed steadily aggraded by about 20 mm in the bed observation area, indicating partial ongoing transport of the incoming sediment. The bed slope gradually increased from aggradation during R2. The large R3 sediment pulse initially caused rapid aggradation and an abrupt increase in slope to near the initial value. After this sediment pulse, the bed aggraded more slowly and slope remained relatively constant, with only minor fluctuations in response to the further, smaller sediment pulses.

After the initial bed winnowing and surface armoring, the GSD of the surface remained remarkably constant (Figure 6c). Surface D_{50} and D_{16} consistently decreased almost immediately in response to sediment pulses (most noticeably the largest pulse of R3), but rapidly recovered back to roughly the same surface size. Surface D_{90} remained relatively unchanged from the supply pulses. Overall, D_{50} of the load remained close to D_{50} of the supply, most consistently during constant feed runs R2 and R6 (Figure 6c). During runs with extended time without supply (most notably, R3 and R7), the load tended to decrease in size, coincident with the decrease in transport rates (Figure 6a).

Figure 6. Time-series of (a) total sediment transport rate, (b) mean bed elevation along the flume (12 m) and the observation reach (2 m), and the bed surface slope along the flume (12 m); (c) D_{16} , D_{50} , and D_{90} of the bed surface (subscript s) and of the bedload at the outlet (subscript l); (d) is the standard deviation (σ) of the bed elevation along the observation reach; (e) is the number (left hand) and percent occupied area (right hand) of potential anchor stones ($>D_{80}$); (f) is the number of clusters (left hand) and the percent area (right hand) developed during the experiment; and (g) cumulative number of new, destroyed, expanded and contracted structures. The sediment transport is plotted for direct comparison with the bed evolution. Purple lines in the top panel indicate timing of sediment pulses, drawn for visualization (not for scale) (modified from Hassan et al., 2020).

Figure 7. An example of bed composition (texture and structure) at the end of run R4 (four sediment pulses). The bottom three pictures show close-up images of the bed surface. The two pictures in the left show bed surface structures, while the right one shows a single large green particle within a patch of fine sediment, which is not classified as a structure. The green, white, and blue stones (22–64 mm) cover 23.3% of the total area.

The standard deviation of local surface elevation (σ in Aberle & Smart, 2003), a measure of overall surface roughness, abruptly increased from ~ 8 mm, with significant fluctuations, to 11 mm at the onset of R3, then declined with generally less significant fluctuations to 9 or 10 mm at the end of the experiment (Figure 6d). The ratio $\sigma/D_{50s} \approx 10/17 \approx 0.6$ through stages R3 to the end.

Figures 6e–6f show the evolution of large grains (that were anchor stones for grain clusters) and grain structures during the overall experiment, which were mapped using consistent criteria (Figure 7). Most mapped structures were clusters or stone cells and very few stone imbrications (Hassan et al., 2020). Isolated large stones were also prominent (Figure 6e). At the very beginning of the experiment, the non-water-worked bed had a small number of random particle groupings that met the structure criteria. The number and area of structures increased rapidly, particularly during R2 with gradual feed constantly modifying the bed surface. However, the large, abrupt sediment pulse in R3 dramatically reworked the bed surface and decreased the number of mapped structures. The four sediment pulses in R4 led to similar but less pronounced adjustments of the bed surface, while the two sediment pulses in R5 (especially the second one) had the expected proportionally greater effect on the bed (Figures 6e–6g). After the sediment pulses (i.e., beginning of R6), the structures were more protruding and stable (Hassan et al., 2020), although they covered a smaller surface area (Figure 6f). The much larger number of prominent individual stones exhibited a similar trend to that of the composite structures until after the major R3 sediment pulse, when it settled to a stable number between 95 and 100, occupying about 10% of the bed area. With constant sediment feed in R6 and no feed in R7, the grain sizes and structure coverage all showed small fluctuations, but remained relatively stable (Figures 6c and 6f), in contrast to R1 and R2.

Insights into the evolution of the aggregate bed structures throughout the experiment is gained by calculating the survival time of each identified structure (Figure 8). The normalized histogram (labeled frequency in Figure 8a) shows that the distribution of survival times was bimodal: $\approx 27\%$ of identified structures were unstable and persisted ≤ 2 h, but many clusters lasted much longer. Based on the cumulative distribution, the median survival time was about 10 h (Figure 8a). About 10% of the structures persisted for 80 h or more. Figures 8b–8e provide two examples (C26 and C50) of structures that persisted for almost 200 h. Both of

Figure 8. (a) Normalized histogram of survival percentage (red, labeled frequency) and corresponding cumulative distribution function (CDF) of structures throughout the experiment, (b and c) structure 26 at 82 h (R3) and 280 h (end of R7), and (d and e) structure 50 at 83 h (R3) and at 280 h (end of R7).

them were formed after the large pulse of R3, but subsequently continued to increase and decrease in size.

4.4. Bed Surface Sediment Exchange

Consistent with clusters exchanging particles, Figure 9 illustrates that the proportion of bed area covered by particles of size larger than 11 mm changed rather significantly during most of the experiment, especially R1–R5. At the same time, the data suggest that the bed adjusted by the end of the experiment such that constant feed R6 and no-feed R7 exhibited smaller changes in surface size classes than R1 and R2 (Figure 9b). In general, coarser size classes showed smaller responses to feed pulses, and tended to slightly decrease on the bed surface immediately after a pulse, or to experience no change (e.g., 22–32, 32–45, 45–64 mm), presumably because finer size classes were deposited on the top of these relatively immobile grains. The change in bed surface mobility in R2 and R6 (constant feed) may reflect the effects of the initial bed surface arrangements inherited from previous runs. Interestingly, even though Figure 9b shows that the size class proportions changed a lot, Figure 6c demonstrates that surface percentiles changed much less. This comparison suggests that grain exchange was widespread (as shown in Figure 9c), but at the same time, increases and decreases at the size class level offset each other, keeping surface percentiles relatively unchanged. It may seem surprising that even when 30%–50% of the bed surface was changing (Figure 6f), surface GSD remained relatively the same (Figure 4c). This means that, although particles on the bed do move, it seems that they do so by exchange of similar sizes so that the bed texture remains relatively constant over time.

4.5. Predicting Grain-Size Ratios From Transport

Equation 9 allows predictions to be made of the ratio of surface D_{50} to load D_{50} if a value for the sediment load threshold of motion is assumed. Using measured τ_b and assuming that $\tau_{cl}^* = 0.045$ (see Section 3.2.2), Figure 10 shows that the bed surface was less coarse than predicted for most of the experiment. Results from R3 to R7 follow a trend almost parallel to the 1:1 line. In these runs, the averaged observed ratio is 4.04 ± 0.94 ,

while the expected ratio is 5.20 ± 1.08 ($\pm$ values are the standard deviation). This difference indicates that the bed surface was 28% finer than expected based on the Church et al. (1998) approach. Conversely, R1 and R2 fall below the line. One explanation is that the assumed constant threshold is inaccurate. A threshold lower than $\tau_{cl}^* = 0.045$ for R1 and R2 would raise these points to the 1:1 line. Conversely, a higher threshold – in particular, 28% higher, giving $\tau_{cl}^* = 0.058$ – would bring R3–R7 down to the 1:1 line. These threshold numbers are broadly consistent with the Wong and Parker (2006) model prediction of thresholds of motion (Figure 4a). Figure 10b shows how observed and predicted grain-size ratios changed over time during the experiment. The trends in R3, R5, and R7 are largely explained by changes in the median size of the load (Figure 6c), as surface D_{50} changed relatively little after the beginning of R1.

5. Discussion

5.1. First Hypothesis: Threshold Change With Surface Coarsening and Structure Development

Difficulties in predicting sediment transport in gravel-bed streams arise from complex interactions between the turbulent flow and the granular boundary, which consists of a wide range of grain sizes, shapes, and bed surface structures. Gravel thresholds of motion typically have a great deal of variability. For example,

Figure 9. (a) Total sediment transport rate is plotted for direct comparison with bed surface coverage. (b) Percentage of bed area occupied by grain sizes larger than 11 mm in the observation area. Values at zero hours reflect the original composition of the sediment mixture. (c) Total percentage of the bed area that changed as calculated by the end of each run.

Lamb et al. (2008) found that compiled flume-based thresholds varied between roughly $\tau_c^* = 0.03$ and 0.1 for a channel slope of up to $\sim 2\%$. In our experiment, we found more than half this range of variability, with τ_c^* varying between ~ 0.035 and 0.075 (Figures 4 and 5). The variability was not random, but systematically changed through time in response to boundary conditions and bed state evolution.

Figure 10. (a) Comparison between expected and observed median grain-size ratio (surface to bedload). (b) Temporal evolution of median grain-size ratios surface to bedload. Lack of water surface long profiles during R1 prevent us from calculating expected median grain-size ratios.

While the average threshold value from our overall experiment matches the average slope dependence from the Lamb et al. (2008) compilation (Figure 4a), the τ_c^* variability we observed is not explained by this average slope dependence. The W&P method suggests an increase in average thresholds of $\sim 50\%$ from the start of the experiment to runs R3–R7, from ~ 0.04 to ~ 0.06 ; at the beginning of R3, there is a remarkable change in τ_c^* from ~ 0.04 to ~ 0.05 (Figure 4a), after which the value of the threshold stress oscillates in response to the other sediment pulses. Our implementation of the Church et al. (1998) method, which assumed an initial $\tau_c^* = 0.045$ for a typical unstructured bed, suggests an increase in threshold of around 33%.

In addition, the history of increasing threshold stress tracks the increase of total stress that resulted from slope and flow depth changes as the bed evolved. As the imposed stress increased, we interpret that the bed surface structure changed to counter the increase of sediment transport that would otherwise occur. In this way, the bed state response acted as a negative feedback, inhibiting transport relative to the magnitude of shear stress increases. Using field monitoring data, Rickenmann (2018) similarly found correlations

between threshold stresses (back-calculated from bedload transport models) and shear stresses during seasonal floods.

More specifically, because the effects of grain size are (theoretically) accounted for in our various threshold analyses, we interpret both the increase in average τ_c^* and also the variability in τ_c^* to reflect the combined effects of surface evolution, such as development of coarse-grain structures, protrusion, and perhaps intergranular friction (Hassan et al., 2020; Yager et al., 2018). The different magnitude of the changes in τ_c^* suggests that pulses in sediment feed have the potential of causing sudden changes in the threshold for motion, while the co-evolution of bed surface texture and structuring and transport variability can cause smaller oscillations around average values. Comparison of threshold values in no-feed runs R1 and R7, and in constant-feed runs R2 and R6, similarly demonstrates the influence that differing initial conditions of bed surface structures and other possible surface effects have on channel bed and transport response to changes in sediment supply.

The threshold of motion is a model-dependent parameter. The correlation between the W&P and W&C models (Figure S3) demonstrates that the overall trends in threshold evolution that we observe are not strongly sensitive to the choice of bedload model. Nonetheless, a model difference is that the W&P thresholds are calculated based on only the D_{50} of the load, while the W&C model (and also the Church et al., 1998 method) uses the median sizes of both the load and the bed surface. One effect is that the W&P threshold immediately increases in response to almost all sediment pulses, including for the first measurement made 1 h after the pulse at the start of R3 (Equation 1; Figures 3 and 4a). In contrast, Figure 4c shows that the W&C threshold (calculated using Equations 4, 6, and 7, without Equation 5) decreases immediately following sediment pulses. The response in Figure 4b (W&C, using Equation 5) also tends to show a decrease in threshold for almost every sediment pulse, although in some cases, this decrease comes after an initial increase (e.g., the first pulse in R4). While the W&C threshold does not drop significantly at the start of R3, the threshold in fact stays low for the first two measurements (2 h), and does not show an increase until 4 h after the pulse. We infer that the bed was still relatively loose at the start of R3, and so the influx of sediment did not make the bed even more mobile. In contrast, for later pulses, the bed was more stable and interlocked prior to each pulse, and the pulse of transported sediment caused W&C model thresholds to initially decrease.

All of the methods used to determine critical shear stress show that thresholds started at low values when the non-water-worked bed was highly mobile, but also remained relatively low during R1 and R2 (Figures 4, 5, and 10). In contrast, the largest sediment supply pulse, at the start of R3, caused a relatively rapid increase in the threshold, suggesting that a combination of deposition and reworking of the preexisting bed surface by increased sediment transport can enhance bed stability. The effect of bed reworking and fluvial deposition by sediment pulse R3 may be comparable to previous work which has found significant differences in bed stability between nonwater-worked and water-worked beds (e.g., Cooper & Tait, 2009; Perret et al., 2018). The R3 sediment pulse caused roughly 10 mm of rapid bed aggradation on average (Figure 6b). Ten millimeters is comparable to $\sim D_{60}$ of the bulk sediment GSD and $\sim D_{25}$ of the average surface GSD (Figures 2 and 6c). While not very different in overall magnitude from the cumulative but gradual aggradation that occurred during R2, the increase in threshold suggests that the short duration of R3 aggradation (and perhaps other surface changes) enhanced bed stability much more, shifting the bed to a different state of higher threshold that was largely maintained, with modest variability, for the rest of the experiment in spite of the supply perturbations.

5.2. Second Hypothesis: Transport Rates Evolve With Armor Development and Surface Structure Development

Perhaps surprisingly, the time series of thresholds of motion is essentially uncorrelated with the degree of aggregate bed structuring (Figures 4 and 6d). The R^2 value obtained by linearly interpolating the number of aggregate structures with the threshold of motion obtained with Equations 9 and 3 was <0.04 . As measured using the method of Hassan et al. (2020), the number of structures began low on the initially screened bed, and increased somewhat during R1 erosion. The number of structures increased most rapidly during the constant feed of R2, even as thresholds remained low. Subsequently, each sediment pulse created a sudden

Figure 11. Temporal rate of change of structure activity. Structure activity includes new, destroyed, expanded, and contracted structures presented in Figure 4d. Run-averaged results are presented as red dashes, while black curves plot instantaneous activity for each run. The instantaneous rate of change was obtained by summing the temporal structure data in Figure 4d, and using central differences.

increase in the number of detected structures, followed by a steady apparent decrease (see Figure 11). The most dramatic sediment pulse (R3) created the extreme structural response.

These results indicate that the bed surface evolved to increase stability even with varying sediment supply regimes and that the development of structures contributed to stabilizing the bed surface. Furthermore, change in the number of newly formed and destroyed structures, and changes to the existing structures, continued through R4–R7 (Figure 6g), indicating continued dynamic response to the flow. Yet the similarity of formed/destroyed and expanded/contracted plots indicates the overall stability (Figure 6g). The small balance of created and expanded features indicates slow additional development within the time-limited experiment.

Hence, the data also demonstrate that clusters remained dynamic throughout the experimental runs, with grains within clusters being exchanged between the bed and bedload (Figure 8). Figure 11 shows that changes in clustering (defined as aggregate structure activity, which combines new, destroyed, expanded, and contracted structures) follows a pattern consistent with changes in sediment supply: a rapid spike occurred right after each change in sediment supply followed by a gradual decline as time passed. The instantaneous rate of change of aggregate bed structures seems to mimic the temporal evolution of bedload transport (Figure 3a). This could be indicating that bed structuring is an inherent process associated with regulating sediment transport and not just part of the physical mechanisms by which the bed surface stabilizes, as previously recognized (Church et al., 1998). In addition, the largest structure activity, caused by the R3 pulse, corresponds to the largest increase in threshold as the bed stabilized. Perhaps, bed structuring is more of an effect of the transport conditions that lead to bed stability (and higher thresholds of motion) than a dominant cause of increased thresholds themselves. The similar temporal behavior of individual large stones, barely mobile as a class, lends support to this conjecture.

If surface structures do not primarily cause increased thresholds of motion, then what does? Bed roughness, which includes the effects of individual particles as well as aggregate structures, might represent a more comprehensive measure of the stability of the bed (e.g., Hassan et al., 2019; Ockelford & Haynes, 2013). Masteller and Finnegan (2017) found that bed stability varied with coarse-grain protrusion. Hodge et al. (2020) used X-ray tomography to measure complete surface grain geometries, and found that protrusion is more important than pocket geometry in controlling thresholds for individual grains. Their measurements included only stones that were free to move (i.e., not constrained by other stones). Hodge et al. (2020) found that clast protrusion (P) was the only one among several measures of the bed that correlated significantly with τ_c^* . Adopting $\sigma = 10$ mm (Figure 6) as an approximation of their measure of protrusion for the post-R3-developed surface, $P/D_{50b} \approx \sigma/D_{50b} = 2.0$, and $\tau_c^* = 0.036$ from the relation of Hodge et al. (2020). This value, derived from the relatively large value of σ , is lower than the values displayed in Figure 4, implying that general surface roughness accounts for only part of the resistance to flow and reluctance of bed particles to move.

Another possible control on critical stress, recently explored by Yager et al. (2018), is that intergranular friction plays a significant role in thresholds, and varies with bed packing and grain interlocking that occur in imbrication and in aggregate structures. Interlocking of grains is also likely to depend on surface and subsurface packing and compaction (e.g., Cooper & Tait, 2009; Marquis & Roy, 2012; Perret et al., 2018). Thresholds and structuring both started low at the beginning of our experiment and increased, but then varied essentially independently. Our data may suggest that once the bed is sufficiently structured and interlocked, thresholds are relatively insensitive to changes in structuring because bed intergranular friction is sufficient to continue stabilizing the bed.

Building on previous work demonstrating correlations between gravel-bed stability and surface roughness statistics (e.g., Cooper & Tait, 2009; Hassan et al., 2019; Masteller & Finnegan, 2017; Ockelford &

Haynes, 2013), Perret et al. (2020) quantified how the statistics of bed surface roughness, over a range of spatial scales from individual grains to bedforms, could improve predictions of bedload transport rate at low Shields stresses. They proposed an empirical relation between spatial statistics and the scaling exponent on shear stress in a bedload transport relation similar to Recking (2012). They found a much weaker dependence of reference stress (their model's threshold parameter) on bed roughness statistics. Nonetheless, the analysis of Perret et al. (2020) suggests that roughness statistics that account for multiple lateral scales of grain interactions may be able to account for grain interlocking and other stabilizing effects. We note that their experiments used a very narrow gravel GSD, such that armoring could not form in their experiments, which may have also inhibited other stabilizing effects related to spatial sorting of grains. In any case, our study explores a complementary parameter space in that we used a constant discharge with sediment supply pulses, while Perret et al. (2020) used a variable discharge hydrograph with no upstream sediment supply.

5.3. Third Hypothesis: The History of Previous Sediment Supply Perturbations Affects Threshold and Channel Response

The initial sediment supply exerted major control on temporal patterns of sediment transport rate and the bed surface evolution (topography and structures). However, bed surface GSD rapidly stabilized (Figure 6) and, after the major R3 sediment pulse, the physical character of the bed stabilized, experiencing only minor perturbations as further sediment pulses passed through the flume.

We conducted the experiment under a transport regime in which particle mobility declined with increasing grain size; the movement of large particles (>22 mm) declined rapidly after the feed and was sporadic (Figure 3)—the regime of partial transport (Wilcock & McArde, 1993) that is typical of gravel-bed streams. About 95% of particles larger than 32 mm (from the bed and in the feed) remained within the flume. The large, mainly immobile “keystones” or “anchor stones” play a major role in stabilizing the bed surface and in the development of bed surface structures, and then in modulating the effect of the subsequent transport.

The areal concentration on the bed of coarse particles (>22 mm, which exceeds D_{80}) that are assumed to trigger the formation of surface structures is in the range 15%–28% (Figure 7b). This range of concentrations is the same as the one shown by Rouse (1965) to maximize flow resistance. Rouse (1965) reported that a coverage of 15%–25% of the bed surface by resistance elements displaces high velocity upward from the bed surface. We did not measure turbulence or near-bed velocities in these experiments, but we note that topographic complexity and protruding coarse-grain clusters in particular enhance turbulence and flow resistance (e.g., Cooper & Tait, 2009; Curran & Tan, 2014). Johnson (2017) also found a correlation between coarse-grain clustering and flow resistance in experimental step-pool channels, where a metric related to Ripley's K function (which measures the degree to which particle spacing deviates from random) predicted the hydraulic flow resistance better than the D_{84} . This suggests that the abundance of coarse particles on the bed, individually and/or arranged into structures, adjusts to maximize flow resistance (Mackenzie & Eaton, 2017), an idea that deserves to be studied in more detail.

Bed surface topography, GSD, and bed surface structures adjusted shortly after the beginning of sediment feed in response to changes in the sediment supply regime. Particularly after structures formed and thresholds of motion indicated that the bed had stabilized, surface GSD was quite insensitive to sediment supply perturbations, and varied less than structures and thresholds of motion. Most changes in the GSD of the bed surface were observed during the non-feed run R1. In subsequent runs, little change in the bed surface GSD was observed before the introduction of the next sediment pulse. Despite repeated sediment pulses, the median size fluctuated around 16 mm (Figure 4c), equivalent to an armor ratio of about 3. Toward the end of the experiment, the median particle size of the bed surface was about the same as that measured at the end of R1 (40 h). The armor ratio persisted regardless of the changes in the sediment supply regime. This outcome is consistent with the assertion of Wilcock and DeTemple (2005) that beds typically remain armored even during relatively high flows, although individual grains within the armor may be mobilized and replaced with transported grains. It seems that, by the end of R1, armoring reached its maximum development for the sediment mixture and the flow conditions that we applied. Any further adjustment of the bed surface to the sediment feed was accommodated by structural development on the bed surface and aggradation throughout the experiment.

Figure 12. Relation between median particle size ratio of the bed surface and the transported material versus the ratio between the shear stress and the critical shear stress associated with the median size of the transported material. All values were calculated for the end of each run. Lines of equal q^* are plotted for comparison.

Figure 12 follows the methodology proposed by Church et al. (1998) and Hassan and Church (2000) to analyze the effect of bed surface coarsening and structuring on sediment transport. For comparison, we plot our results with field data from Harris Creek and Oak Creek, both of which had armored and structured channel beds (Figure 12). Data from Harris Creek and Oak Creek plot around the $q^* = 0$ line (Equation 8), implying that sediment was moving near the threshold condition and that sediment transport rates were low. For our experiment, data from R1 and R2 plot close to $q^* = 0$, whereas the rest of the pulse, feed, and no-feed runs, plot around $q^* = 0.1$. Comparing the no-feed runs, the D_{50s}/D_{50l} ratio for R7 was about 33% larger than that obtained for R1, because structures had saturated the bed in the later run. Similar results were obtained for the constant feed runs; the D_{50s}/D_{50l} ratio was about 27% larger in R6 than in R2, after structures had saturated the bed.

5.4. Consequences: Considerations for Channel Stability

Alluvial streams adjust channel form to accommodate the water and sediment loads introduced into the channel. These adjustments include bed surface grain size, structures and roughness, channel width (hence also depth), channel macroform elements (riffles, pools, bars, and meander form), and net aggradation/degradation. The channel-scale morphology often appears to remain stable over long periods of time even as flood discharge and corresponding sediment supply are variable through time. What mechanisms lead to this stability?

In our experiment, flow remained constant while sediment load varied.

The experimental design meant that channel adjustments were limited to slope, bed surface grain size, surface structures, and other surface and flow characteristics that we did not directly measure (such as intergranular friction and near-bed turbulence). In natural rivers, the volume of sediment passed through an alluvial reach in a short time period tends to be small in comparison with the volume of sediment stored and slope adjustment tends to be slow, although this can vary for individual reaches during larger floods. In our flume, the volume of sediment introduced was relatively large, driving net aggradation and slope adjustment. Over the entire experiment (R1–R7), 1,500 kg of sediment was supplied at the upstream end (R2–R6). Slightly more than 400 kg of that sediment became stored in the channel at the end of R7, or about 30% of the supply (Elgueta-Astaburuaga & Hassan, 2019). The bed surface condition also adjusted. We observed that the GSD of surface grains rapidly assumed a value that remained largely unchanged throughout the sequence of sediment pulses and zero feed (Figure 6c), while the number and location of recognizable grain structures appreciably changed (Figure 6f). Eaton and Church (2009) demonstrated that, in the presence of non-erodible banks, bed state can accommodate changes in sediment supply of up to $4\times$ – the range in pulse magnitudes introduced in the present experiment. Evidently, the main bed stabilizing effect is due to variation in the sediment structures and potentially other unmeasured bed characteristics, at least under steady flow.

Our experiment held water discharge – a principal governing conditions of alluvial channel form – constant while we varied bed material influx, the other principal condition. Over a range in sediment supply, the channel reached and maintained a steady state by adjustment of bed surface conditions (a result observed, however, subject to fixed bank conditions and steady, slow aggradation under conditions of partial transport). As mentioned above, Perret et al. (2018, 2020) report similar and complementary experiments focused on changing discharge in the absence of sediment supply. Another complementary experiment was conducted numerically by Parker et al. (2008) in which hydrographs were cycled through a simple, but long, channel with constant bedload influx. After some distance, bed elevation and bed surface GSD stabilized so that there was no net change from flood to flood, even though individual floods of variable magnitude created transient changes in bedload magnitude and caliber. Considered together, the two effects identify

powerful mechanisms residing in the adjustment of the sedimentary bed surface to withstand a substantial range of flows and bed material inflows, and thereby lend stability to gravel-bed river channels. In the regime of partial transport, channel-scale stability can be obtained in gravel-bed streams by subchannel-scale sedimentary adjustments mediated through sediment transport.

Upland stream channels tend toward bed stability that is maintained through some range of fluctuations in flow and sediment supply regimes by bed state adjustments involving bed structures and potentially other aspects of the bed surface. Although most of the time flows are well below the critical value needed for significant entrainment and transport of sediment, higher flows are likely to change the bed surface and create a transient situation in which the bed develops a new stable state. When the water discharge remains constant, the degree of the bed surface structuring and coarsening depends on the recent history in magnitude and frequency of sediment supply and mobilizing events. Our results show that a simple functional relation between sediment transport rate and shear stress (or any other flow measure) derived on the basis of a constant critical value is unlikely to cover all the phenomena that we documented in our experiment and other similar experiments (e.g., Church et al., 1998; Hassan & Church, 2000; Tait & Willetts, 1992).

6. Conclusions

We conducted a flume experiment to study the impact of changes in sediment supply on channel stability in gravel-bed streams, focusing on surface GSD, bed structuring, and particle thresholds of motion. The bed surface armored rapidly within the first few hours of the 280-h experiment, and the armoring remained intact throughout the rest of the experiment. In response to pulses of sediment feed, the bed surface fined for a short period of time, but then rapidly returned to essentially the same surface GSD. Surface structures developed around larger anchor stones, and individual clusters remained dynamic throughout the experiment by exchanging particles through entrainment and deposition.

We calculated thresholds of motion that varied in time, based on experimental transport rates and shear stresses, using two bedload transport models and a grain-size-based method from Church et al. (1998). The different approaches yielded very similar trends, with the threshold for motion responding to changes in sediment supply and overall increasing during the experiment. We propose that this is due to the increase in bed stability provided by bed structure, since other relevant variables either remained constant (e.g., flow rate), changed very little (e.g., bed surface GSD) or their variability could not explain the observed variability in the critical Shields stress (e.g., channel slope). All three of our specific hypotheses were confirmed, and we thereby demonstrate that modulation of sediment transport by channel bed adjustment to variable flow and sediment influx is an important stabilizing mechanism in gravel-bed rivers.

Our results show that sediment transport responded clearly to pulse changes in sediment supply but the increase did not last long. Once surface structures formed, the estimated critical shear stress was between 17% and 36% higher than the nominal critical Shields number value of ~ 0.045 for widely graded gravels. The value of the Shields number increased by 37% between R1 and R7, while roughly 50% increase was observed between the constant feed runs R2 and R6. Throughout the experiment, bed surface stability increased and bed particle mobility of the larger fraction decreased. Particle mobility changed in response to pulses but returned to a stable value toward the end of the experiment. In conclusion, our experiment showed how a poorly sorted gravel-bed surface responded to changes in sediment supply. Despite little change in the GSD, surface structures constantly formed and evolved throughout the experiment, contributing to stabilizing the bed, increasing the threshold for motion and reducing particle mobility.

Data Availability Statement

Experimental data can be found at <http://doi.org/10.5281/zenodo.1227423>; images at <http://doi.org/10.5281/zenodo.3558356>; digital elevation models at <https://zenodo.org/record/1478895#.W-ICWPZFxuk>, and in the supporting information.

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