Experimental analysis of river bed rock alluviation in different sedimentation and hydraulic environment

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The Pothole in the bed of Denwa River as a result,incision driven by the collision of bedload grains and bedrock requires transport at below capacity condition

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Abstract

[1] Channel incision into bedrock plays a critical role in mountain landscape evolution. A bed that is completely alluviated cannot undergo incision. As a result, incision driven by the collision of bedload grains and

bedrock requires transport at below-capacity conditions (for example, where the bed is partially covered by alluvium). In a mechanistic model of bedrock incision, it is important to have a formula describing the relationship between the bedload transport rate and the degree to which the bed is covered by sediment. The purpose of this work is to understand the mechanisms underlying the degree of bed exposure in bedrock rivers by means of an experimental study. A number of experiments with various hydraulic and morphological bed conditions were performed in order to characterize this model. Here we focus on planar bedrock beds with some roughness. The results suggest that the sediment supply, channel slope, hydraulic bed roughness, the degree of areal coverage, and thickness of antecedent gravel in the channel, as well as the presence of boulders in the channel are major factors controlling bedrock exposure. For sufficiently low ratios of sediment supply to transport capacity, it was found that bedrock roughness can play a role in determining the degree of bedrock exposure. For higher sediment ratios, on the other hand, the composite roughness associated with grain roughness, bars, and/or antidunes dominates, so that the underlying bedrock roughness no longer affects the degree of exposure of the bed. For lower bed slopes (i.e., less than 0.015, based on our experimental setting), bedrock exposure decreases more or less linearly with increasing values of the ratio of sediment supply rate to capacity rate. For higher bed slopes (i.e., more than 0.015), there is a range of lower values of the ratio of sediment supply rate to capacity rate where a bedrock surface becomes fully exposed without any alluvial deposit. For a given bedrock roughness, this range expands solely with increasing slope regardless of shear stress. Within the upper range of values of the ratio of sediment supply to capacity rate a linear relationship between the degree of bedrock exposure and this supply to capacity ratio still prevails. The addition of model boulders (can be viewed as very high hydraulic bed roughness) to the channel has been found to suppress the overexposure of bedrock, and so restore a linear relation between the degree of exposure and the ratio of sediment supply rate to capacity rate. Formulations for estimating bedrock exposure as a function of sediment supply to capacity ratio under different river channel characteristics are proposed. A linear relation between the degree of alluvial cover and the ratio of sediment supply rate to capacity rate is a previously proposed model. The present study expands the result for other cases including runaway alluviation. Some landscape evolution models assume an abrupt shift between fully exposed bedrock and complete alluviation, whereas others assume the linear relation. The present study shows that both formulations are valid under different settings of river characteristics.

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1. Introduction

[2] During the last decade, there has been a rapid advancement in the development of models of mountain landscape evolution. These models have contributed to the understanding of the fundamental mechanisms underlying landscape evolution [e.g., *Howard*, 1994; *Tucker and Slingerland*, 1994; *Whipple and Tucker*, 2002; *Sklar and Dietrich*, 2004; *Gasparini et al.*, 2006; *Lamb et al.*, 2008]. Bedrock incision processes are believed to play important roles in the evolution of Mountain Rivers and their associated hillslopes. When studying incision processes mechanistically, one must be able to answer the following questions: (1) how does bedrock exposure on the river bed relate to the incision rate? (2) How does exposure vary with parameters such as sediment supply and hydraulic conditions? Here only the latter topic is explored. Although several flume experimental studies [e.g., *Wohl and Ikeda*, 1997; *Finnegan et al.*,

2007; *Johnson and Whipple*, 2007] have investigated the feedbacks among sediment transport, bed alluviation, channel incision, and bed morphology on erodible beds, they have not focused on the degree of bedrock channel alluviation under systematic variations of sediment supply and hydraulic conditions, which could guide us toward the modeling of bed cover mechanism in order to use in landscape evolution models.

[3] While there are a variety of mechanisms that may play roles in bedrock incision, including abrasion by suspended load, plucking, and cavitation [*Whipple et al.*, 2000], this study is limited to the case of bedrock incision in those gravel-bed rivers for which the process is driven by gravel striking the bedrock surface. Intuitively, the bedrock incision rate should depend on the degree to which the bedrock surface is exposed to the gravel in transport [e.g., *Sklar and Dietrich*, 1998, 2004]. Boulders entering the channel from basaltic, steep canyon walls have been found to shut off the incision process over long, geological time scales [*Seidl et al.*, 1994], while some bedrock channels with thin alluvial covers in a badlands setting have been found to show significant erosion in as little as seven years [e.g., *Howard and Kerby*, 1983]. Few studies [e.g., *Sklar and Dietrich*, 2004; *Montgomery et al.*, 1996] suggested that for a given drainage area, a broad range in the degree of alluviation was found within a relatively modest range of channel slopes and that the degree of bedrock exposure in mountain streams may be quite dynamic. However relatively few relations supported by experimental data exist to predict the degree of bedrock exposure. Such paucity limits both predictions of bedrock incision and the development of associated landscape evolution models.

[4] Beaumont et al. [1992] were among the first to consider the cover effect (according to which a higher sediment supply rate results in a less incision rate) in bedrock incision. They scaled the incision rate with the degree to which the transport capacity is in excess of the sediment supply. They did not however include the effect of bedload tools [e.g., Sklar and Dietrich, 2004], according to which higher sediment supply can (but not necessarily always does) lead to a greater incision rate. Howard [1994] and Tucker and Slingerland [1994] considered the cover effect in their landscape evolution model as an abrupt shift from the bedrock to the alluvial state, which is assumed to occur whenever the sediment supply exceeds the transport capacity. Slingerland et al. [1997] proposed that both tool effects and cover effects should be important in natural bedrock rivers by using existing studies of slurry pipeline abrasion, such as that of de Bree et al. [1982].

[5] *Sklar and Dietrich* [1998, 2004] have contributed to this perspective by including both cover and tool effects in their mechanistically based saltation–abrasion model. To implement this they assumed a relationship such that the areal fraction of bedrock exposure (*po*) decreases linearly with the ratio of sediment supply per unit width (*qs*) to transport capacity rate (*qc*) for the former as follows.

$$p_o = 1 - \frac{q_s}{q_c} \tag{1}$$

This is the basic formula that is currently widely used in many bedrock erosion modeling studies [e.g., *Gasparini et al.*, 2006; *Crosby et al.*, 2007; *Lamb et al.*, 2007, 2008] and that is being evaluated in this paper. *Sklar and Dietrich* [2004] found that maximum erosion rates occur at a moderate sediment supply ratio because of the tradeoff between the partially alluvial coverage and the availability of the abrasive tools.

[6] It is of value to realize that the assumption of *Sklar and Dietrich* [1998, 2004] is the simplest reasonable way to characterize the effect of variable cover. It thus represents the obvious first assumption for treating the problem and at the same time opens the way for verification and/or improvement of the cover relation. *Sklar and Dietrich* [2002] and *Sklar* [2003] have obtained support for the linear relationship between the bedrock exposure and the sediment supply ratio through the use of flume experiments. They varied the sediment supply of uniform-size gravel while holding slope and water discharge constant and observed the resulting formation of sediment patches over a smooth surface. As the sediment supply rate was increased systematically, they observed that the degree of cover increased and that grain collisions became more frequent. Beyond a "threshold of alluvial deposition", they found that the number and size of the depositional patches grew in proportion to the increase in supply rate. They concluded that these observations confirmed their linear relationship for bed cover.

[7] *Demeter et al.* [2005] also investigated the influence of bed roughness on partial alluviation by using an experimental flume. They found that low-roughness beds require a relatively high sediment supply before any alluvial patches form. As the sediment supply increases, they found that the channel can accommodate only low levels of partial coverage before a "runaway alluviation" rapidly converts the bed to an aggrading alluvial bed. In contrast, they found that a rougher bedrock bed partially alluviates at lower sediment supply rate and does not experience runaway alluviation at higher supply rates.

[8] These valuable experiments have helped motivate the present work. This notwithstanding, these experiments were limited to: (1) very low channel slope (≤0.007), almost the lowest end of slopes in natural bedrock rivers (0.001–0.2) [e.g., *Sklar and Dietrich*, 1998; *Montgomery and Buffington*, 1997; *Wohl and Merritt*, 2001]; (2) very low width to depth ratios (≤3) in order to suppress alternate bar formation despite the fact that alternate bars (or pool-riffle) are commonly found in bedrock rivers with slope = 0.001–0.03 [e.g., *Montgomery and Buffington*, 1997; *Wohl and Merritt*, 2001]; and (3) a bedrock configuration with relatively low roughness (and no addition of immobile "boulders"). *Demeter et al.* [2005] also represent a continuation of the work of *Sklar and Dietrich* [2002], in that they examined particle collisions associated with an increase in sediment supply rate in a channel starting from a bare bedrock surface. However their experiments were not able to answer the following question; how would the channel response in terms of bed exposure if the starting point were a bed with pre-existing random alluvial patches?

2. Overview of the Present Work

[9] The experiments reported here differ from previous work in three ways. Firstly, the experiments were commenced with at least some antecedent cover of alluvium. Secondly, the range of slopes studied was significantly higher than earlier experiments that studied the degree of bedrock exposure as a function of sediment supply to capacity ratio (qs/qc) [e.g., *Sklar and Dietrich*, 2002; *Demeter et al.*, 2005], and in fact the slopes studied here are within the range of field bedrock streams. Thirdly, the width-depth ratios of the flow were much larger (between 11 and 31) than previous experiments, and thus allowed for the formation of bars. These three factors help make the experiments reported here better models of actual field conditions.

[10] Random patches of sediment coming from hillslope failures or tributaries are prevalent along the banks in natural bedrock streams. Some patches are large and even cover the majority of the bed. The condition of abundant sediment supply from hillslopes is commonly observed in actively uplifting areas such as the Coast Range of California, USA, from our field survey. It is less common in areas where the rates of weathering and sediment supply to the

channel are low, such as Piccaninny Creek, Australia [e.g., *Wohl*, 1993; shown below]. As shown here, commencing the experiments with randomly placed alluvial patches can result in results that differ from experiments started from a bare bedrock surface. In low-slope channels (for example, lower than 0.01, as observed in our flume experiments) moving grains typically collide in such a way as to result in the formation of patches. In high-slope channels (for example, slopes in excess 0.01) however even when the bedrock surface itself is rather rough, such patch formation may not be observed.

[11] Three types of bedrock surfaces were employed in the experiments reported here. In all of them however the roughness of the surface irregularities of the bedrock itself, as determined from a Manning–Strickler resistance relation (described below), was less than that for their partially alluviated or fully alluvial counterparts. As a result, the unalluviated bedrock surfaces were able to accommodate much higher bedload transport rates without alluviation (in high-slope channels in excess 0.01). When an antecedent alluvial cover (patches of sediment or uniformly continuous cover) was added however further alluviation was observed at much lower transport rates. The alluvium in partially alluviated channel tends to cause particles in transport to deposit. This effect is due to, among other things, the composite roughness height of the bedrock surface, alluviated grains and bedforms such as bars and antidunes. These factors result in a lower overall flow velocity and a tendency for colliding particles to come to rest and accumulate.

[12] As hypothesized by *Sklar and Dietrich* [1998, 2004], the ratio *qs*/*qc* is assumed to have a controlling effect on the degree of alluvial cover over a bedrock surface. A physical model for studying the variation of bedrock exposure with the sediment ratio *qs*/*qc* can play a valuable role in delineating the physical mechanisms underlying bedrock exposure, and thus allow for the development of a more realistic and advanced model of bedrock incision processes. The aims of the present research are: (1) to investigate the fundamental mechanisms controlling the degree of bedrock exposure in steep mountain streams; (2) to examine the validity of the simple formulas used in previous landscape evolution models [e.g., *Howard*, 1994; *Tucker and Slingerland*, 1994; *Whipple and Tucker*, 2002; *Sklar and Dietrich*, 2004; *Lamb et al.*, 2007, 2008]; and (3) to formulate a more advanced model of the cover/exposure effect by using results from an experimental flume.

3. Experimental Procedure

[13] The experiments were conducted in the Jaypee HEW Laboratory, Rewa. This study is based on the experimental study of Phairot Chatanantavet and Gray Prker(2008) The metallic-walled tilting flume has a width *B* of 0.9 m, a length of 13 m long and a bed slope that can be adjusted up to 0.053. Water discharge can be read from a manometer with a maximum value of 250 l/s. Sediment feed is discharged from two sediment feeders at the upstream end with a combined rate of up to 370 g/s. The sediment transport is circulated back to the feeders by a recirculating system by using a jet pump. Two essentially uniform sediment grain sizes of 2 mm and 7 mm were employed for the experiments. The two sizes were not mixed. The depth of the alluvial deposit was measured using a probe that can penetrate sediment patches down to the "bedrock" surface and the gravel depth can be read directly from the built-in scale. Aerial photographs were taken with a digital camera attached to a movable cart.

[14] A non-erodible bedrock surface was fabricated using a mixture of sand, cement, and vermiculite. Figure 1 shows the hand-made bedrock configurations consisting of longitudinal grooves (LG), random abrasion type I (RA1), and random abrasion type II (RA2). A field example of a bedrock bed with longitudinal grooves is illustrated in Figure 2, which is from the Denwa River Rift valley of Pachmarhi, India.

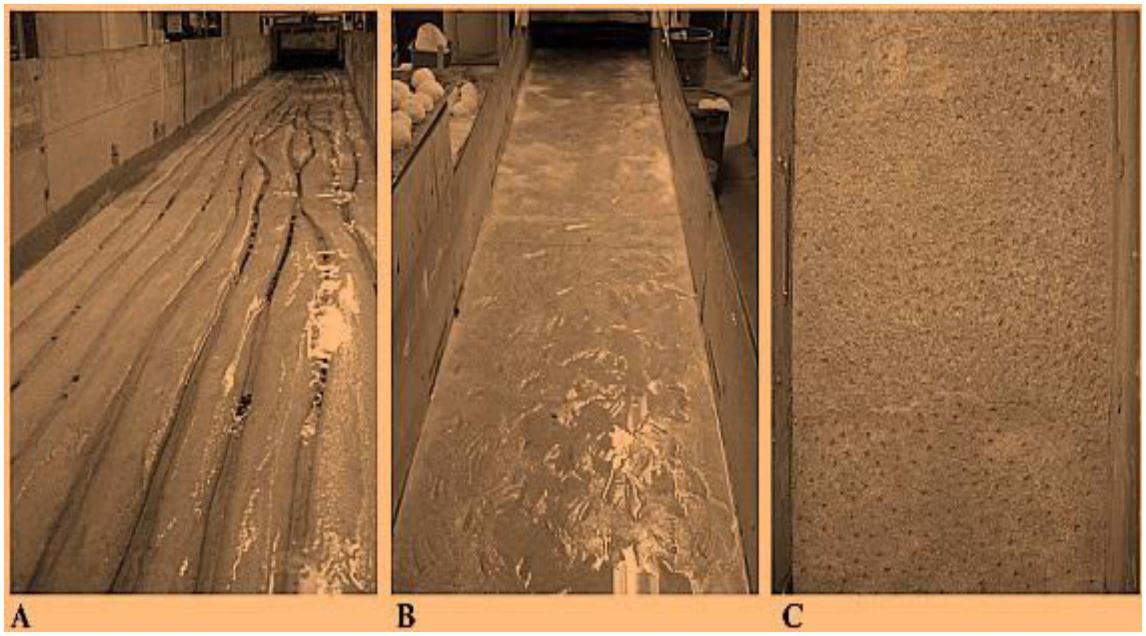

Figure 1.Simulated bedrock morphologies. From left to right: (a) longitudinal grooves (LG); (b) random abrasion type 1—smooth (RA1); and (c) random abrasion type 2—rougher (RA2).

Figure.2 An example of the field condition of bedrock bed with longitudinal grooves: Denwa Rift Valley of Pachmarhi, India.

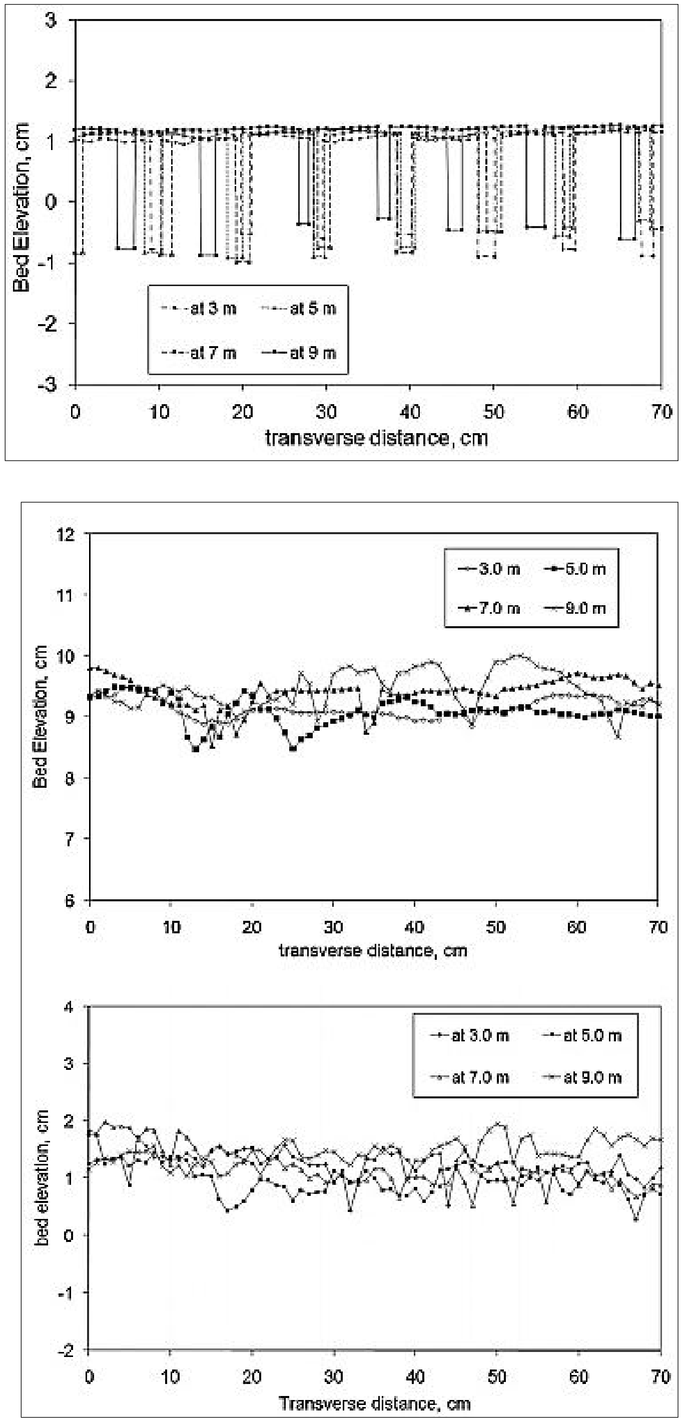

Figure 3. Measured transverse topographies of the model bedrock bed for LG, RA1, and RA2, respectively, for several transects from upstream to downstream, referenced from the upstream end of the flume. For reference, the outlet of the feeder was located 1.5 m downstream of the upstream end of the flume. Longitudinally averaged standard deviations of bed elevation fluctuations for LG, RA1, and RA2 are 6.7 mm, 2.4 mm, and 2.7 mm, respectively. The hydraulic roughness height for LG, RA1, and RA2 are 0.4 mm, 0.4 mm, and 3.5 mm, respectively. Note that the flume width is 0.9 m. Only bed elevation transects corresponding to the middle 0.7 m of the flume are shown here. The flume length is 13 m.

[15] The hydraulic roughness height k_{sb} of each bedrock surface type (LG, RA1, and RA2) was back-calculated from the Manning–Strickler relation, which is defined as follows.

$$\frac{U}{u_*} = a_r \left(\frac{H}{k_{sb}}\right)^{1/6} \tag{2}$$

where H denotes the mean flow depth, U denotes the mean flow velocity, u_* denotes the shear velocity, given by the relation

$$u_* = \sqrt{gHS} \tag{3}$$

where g denotes the acceleration of gravity, S denotes bed slope, and a_r is a constant. *Parker* [1991] proposed a value for αr of 8.1 for gravel-bed rivers based on field data. This value has been confirmed by *Wong and Parker* [2006] and *Wong et al.* [2008] in flume experiments. The latter paper is particularly relevant to the present study, as the sediment material used in that study is identical to the coarser sediment (7 mm) used here. It should be noted here however that here k_{sb} refers to the intrinsic hydraulic roughness of the bedrock surface and that the measurements of flume-averaged hydraulics for the back-calculations of k_{sb} were done without any sediment present.

[16] Figure 3 provides transects of bed elevation of the three bedrock types, illustrating the differences in type and degree of hydraulic roughness. The standard deviations for bed elevation fluctuations for the LG, RA1, and RA2 beds were found to be 6.7 mm, 2.4 mm, and 2.7 mm respectively. The corresponding computed hydraulic roughness heights from flow measurements for LG, RA1, and RA2, as back-calculated from flow measurements and a Manning–Strickler resistance relation are 0.4 mm, 0.4 mm, and 3.5 mm respectively. Note that although standard deviations for RA1 and RA2 are close, their hydraulic roughness heights differ nine-fold. In contrast, although standard deviations for LG and RA1 differ by a factor of almost three, their hydraulic roughness heights are the same. Note that *Johnson and Whipple* [2007] and *Finnegan et al.* [2007] have used the standard deviation of bed topography as their measures of bed roughness.

[17] For slope S = 0.03, 20-cm model boulders were also added to bed configuration RA1 to investigate the effect of boulders (which significantly increase hydraulic roughness) on bedrock exposure. The configuration of model boulders is illustrated in Figure 4. Figure 5 shows an example of a field stream, i.e., the Sonbhadra River in Pachmarhi, India, which has large, probably colluvial boulders that partially cover the bed.

[18] Each experimental run was performed using specified values of bedrock surface slope, water discharge, and sediment supply, and continued until a steady state with relatively no changes in bedrock exposure was achieved. Such a state required between 3 and 30 hours of run time. Lower slopes required longer times for the system to adjust morphologically to a steady state. Each experiment was run separately; in no case did the starting condition of one run consist of the final condition of a previous run.

Attention was paid to make sure that the sediment mass per unit area was more or less uniform all over the channel during the runs by using the probe measuring the gravel depth. In most runs, unless otherwise noted, the experiments were commenced with either hand-placed gravel patches (at lower slopes, for example, less than 0.015) or a continuous layer of gravel with uniform thickness of about 2–5 cm thick in the channel, so as to simulate sediment supply from small landslides.

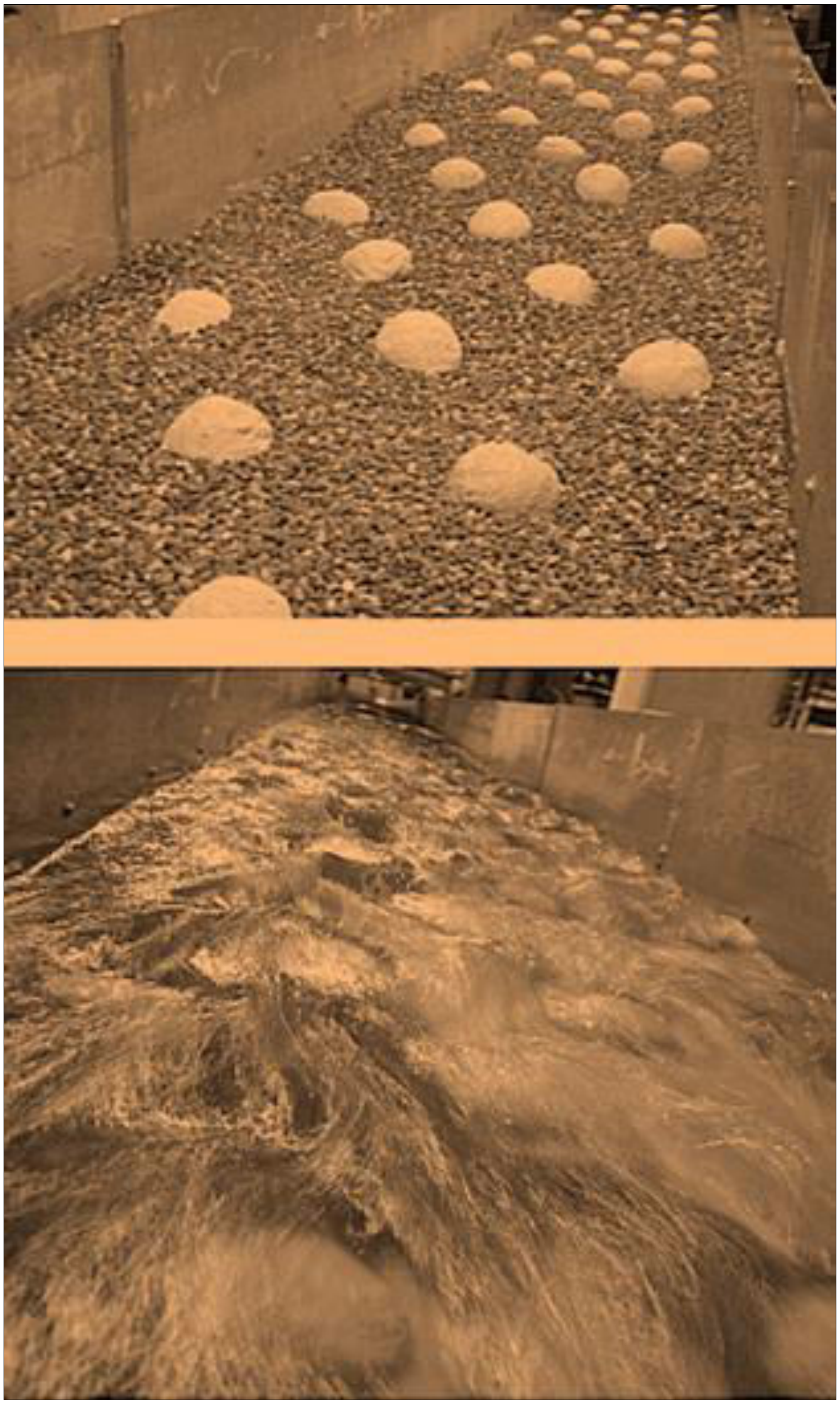

Figure 4.Experimental flume after adding model boulders: (a) before commencing the flow, and (b) after commencing the flow. The runs in this configuration are Runs 4E-1 to 4E-6.

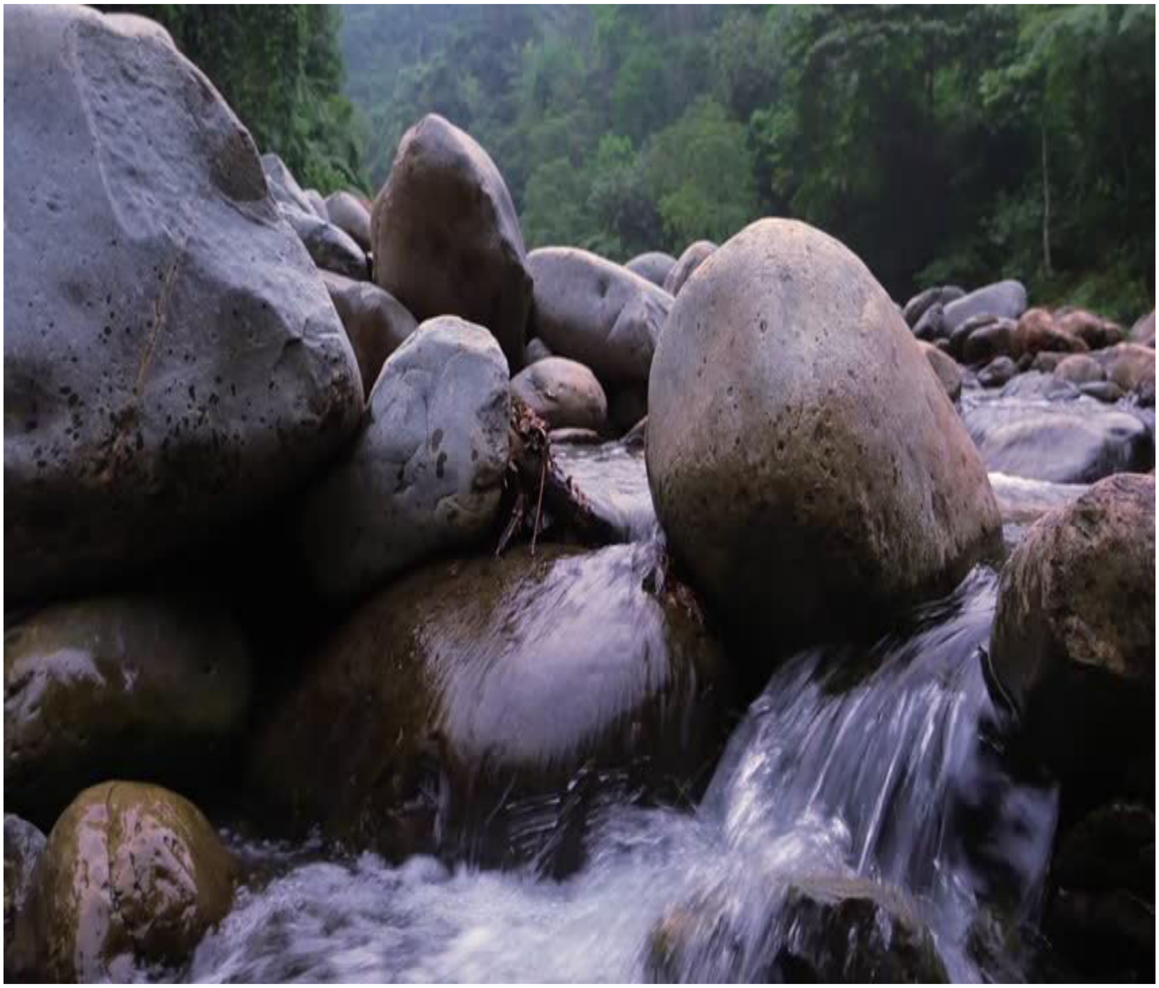

Figure 5. An example of a field river with large boulders in the stream bed: Sonbhadra River, Pachmarhi ,India. Note the bar formation between the boulders.

These antecedent patches are illustrated in Figure 6. During each run, aerial photographs were taken every 15–60 minutes depending on the slope used. The images were taken while the water was flowing. The photographs were then processed using the program ImageJ ® to compute the areal fraction of bedrock exposure.

4. Results

4.1. Summary of Results

[19] The experimental results show that interactions among the sediment supply, channel slope, hydraulic bed roughness of either bedrock surface (in case of flow over bare bedrock) or mixed alluvial-bedrock surface (for example, bars or dunes), and grain size play essential roles in controlling the overall degree of bedrock exposure. Note that unless stated otherwise in this section, the results are obtained under the assumption that there is an abundant supply of sediment coming to the channel from hillslopes.

Figure 6. An example of gravel patches placed before commencing runs simulating an initial state with sediment from landslides. This configuration was used for all runs with slopes less than 0.015, except series 2-Ax and 2-Dx-a to e, which began from a bare bed. For all runs with slopes greater than or equal to 0.015, the runs were commenced with a complete alluvial cover with uniform thickness ranging from 1 to 6 cm. For Run 2-Bx, the run began from a bare bed.

The differences among Qc, Qal,c, and Qo,c (or the values per unit width qc, qal,c, and qo,c) can be explained as follows. Qc is a generic term used here to refer to a capacity sediment transport rate in general. However, Qal,c is a term mostly referred to in this paper and was determined experimentally by starting a run with some antecedent alluvium and increasing the sediment supply until the bed was completely covered by alluvium. Qo,c, on the other hand, is rarely referred to in the results section (yet important to distinguish them) and was determined experimentally by starting a run with a bare bedrock bed and increasing the sediment supply until the bed was completely covered by alluvium and followed by aggradation. The details regarding these capacity transport rates are also discussed more in the sections 4.2 and 4.3.

Table 1. Measured Flow, Bed Type, and Sediment Transport Conditions in Main Experimental Runs

	Bed type	mm		cm	l/s	g/s	g/s			mm	cm	cm/s				Hours
Run		k_{sb}	S	h_{si}	Q_w	Q_s	Q_c	q_s/q_c	p_0	D	H	U	**Fr**	τ^*	*B/H*	t_r
1-A1	LG	0.4	0.0115	1.5	24	2.6	25*	0.10*	0.77	2	3.5 ± 1.0	76	1.3	0.12	26	3.0
1-A2	LG	0.4	0.0115	1.5	24	7.2	25*	0.29*	0.50	2	3.5 ± 1.0	76	1.3	0.12	26	5.25
1-A3	LG	0.4	0.0115	1.5	24	13.6	25*	0.54*	0.39	2	3.5 ± 1.0	76	1.3	0.12	26	9.00
1-A4	LG	0.4	0.0115	1.5	24	20.0	25*	0.80*	0.17	2	3.5 ± 1.0	76	1.3	0.12	26	7.17
1-A5	LG	0.4	0.0115	1.5	24	25	25*	1.0*	0	2	3.6	67	1.12	0.125	25	6.0
1-B1	LG	0.4	0.02	2.0	55	40.7	110*	0.37*	1	7	5.0	122	1.74	0.09	18	3.5
1-B2	LG	0.4	0.02	2.0	55	62	110*	0.56*	0.42	7	5.5 ± 1.5	111	1.51	0.11	16	11.75
1-B3	LG	0.4	0.02	2.0	55	83	110*	0.75*	0.27	7	5.5 ± 1.5	111	1.51	0.11	16	25.0
1-B4	LG	0.4	0.02	2.0	55	110	110*	1*	0	7	6.0	102	1.32	0.095	15	5.0
2-C1	RA1	0.4	0.016	1.5	55	17.4	87*	0.20*	1	7	6.0	102	1.33	0.083	15	3.0
2-C2	RA1	0.4	0.016	1.5	55	23.7	87*	0.27*	0.61	7	7.0 ± 2.0	87	1.05	0.097	13	3.25
2-C3	RA1	0.4	0.016	1.5	55	36.7	87*	0.42*	0.49	7	7.0 ± 2.0	87	1.05	0.097	13	6.0
2-C4	RA1	0.4	0.016	1.5	55	87	87*	1*	0	7	7.0	87	1.05	0.097	13	6.25
2-E1	RA1	0.4	0.03	3.0	55	140	264*	0.53*	1	7	4.0 ± 1.5	153	2.4	0.10	22	1.0
2-E2	RA1	0.4	0.03	3.0	55	185	264*	0.7*	0.23	7	4.0 ± 1.5	153	2.4	0.10	22	1.0
2-E3	RA1	0.4	0.03	3.0	55	264	264*	1*	0	7	4.0 ± 1.5	153	2.4	0.10	22	0.75
3-C1	RA2	3.5	0.016	1.0	40	25	125*	0.20*	1	2	4.5 ± 1.0	99	1.49	0.21	20	2.0
3-C2	RA2	3.5	0.016	1.0	40	35	125*	0.28*	0.58	2	5.0 ± 1.0	89	1.27	0.22	18	1.75
3-C3	RA2	3.5	0.016	1.0	40	43.7	125*	0.35*	0.42	2	5.0 ± 1.0	89	1.27	0.22	18	1.75
3-C4	RA2	3.5	0.016	1.0	40	71.5	125*	0.57*	0.35	2	5.0 ± 1.0	89	1.27	0.22	18	2.75
3-C5	RA2	3.5	0.016	1.0	40	97.6	125*	0.78*	0.20	2	5.0 ± 1.0	89	1.27	0.22	18	1.2
3-C6	RA2	3.5	0.016	1.5	55	17.4	87*	0.20*	1	7	6.5	94	1.17	0.09	14	2.0
3-C7	RA2	3.5	0.016	1.5	55	23.7	87*	0.27*	0.61	7	7.0 ± 2.0	87	1.05	0.097	13	3.25
3-C8	RA2	3.5	0.016	1.5	55	36.7	87*	0.42*	0.49	7	7.0 ± 2.0	87	1.05	0.097	13	6.0
3-C9	RA2	3.5	0.016	1.5	55	87	87*	1*	0	7	7.0	87	1.05	0.097	13	6.25
2-A1	RA1	0.4	0.0115	1.5	24	2.6	25*	0.10*	0.86	2	3.5 ± 1.0	76	1.3	0.12	26	3.5
2-A2	RA1	0.4	0.0115	1.5	24	7.2	25*	0.29*	0.76	2	3.5 ± 1.0	76	1.3	0.12	26	7.0
2-A3	RA1	0.4	0.0115	1.5	24	13.6	25*	0.54*	0.41	2	3.5 ± 1.0	76	1.3	0.12	26	8.0
2-A4	RA1	0.4	0.0115	1.5	24	20.0	25*	0.80*	0.19	2	3.5 ± 1.0	76	1.3	0.12	26	10.0
2-A5	RA1	0.4	0.0115	1.5	24	25	25*	1.0*	0	2	3.6	74	1.25	0.13	25	3.0
2-B1	RA1	0.4	0.02	2.0	55	40.7	110*	0.37*	1	7	5.0	122	1.74	0.09	18	3.0
2-B2	RA1	0.4	0.02	2.0	55	62	110*	0.56*	0.41	7	5.5 ± 1.5	111	1.51	0.11	16	5.0
2-B3	RA1	0.4	0.02	2.0	55	83	110*	0.75*	0.29	7	5.5 ± 1.5	111	1.51	0.11	16	5.0
2-B4	RA1	0.4	0.02	2.0	55	97	110*	0.88*	0.22	7	5.5 ± 1.5	111	1.51	0.11	16	3.0
2-B5	RA1	0.4	0.02	2.0	55	110	110*	1*	0	7	6.0	102	1.32	0.095	15	5.0
3-A1	RA2	3.5	0.0115	1.5	24	2.6	25*	0.10*	0.76	2	3.5 ± 1.0	76	1.3	0.12	26	4.0
3-A2	RA2	3.5	0.0115	1.5	24	7.2	25*	0.29*	0.45	2	3.5 ± 1.0	76	1.3	0.12	26	8.5
3-A3	RA2	3.5	0.0115	1.5	24	13.6	25*	0.54*	0.39	2	3.5 ± 1.0	76	1.3	0.12	26	8.0
3-A4	RA2	3.5	0.0115	1.5	24	20.0	25*	0.80*	0.16	2	3.5 ± 1.0	76	1.3	0.12	26	7.0
3-A5	RA2	3.5	0.0115	1.5	24	25	25*	1.0*	0	2	3.6	74	1.25	0.13	25	3.0
3-B1	RA2	3.5	0.02	2.0	55	40.7	110*	0.37*	1	7	5.5 ± 1.5	111	1.51	0.11	16	1.0
3-B2	RA2	3.5	0.02	2.0	55	62	110*	0.56*	0.39	7	5.5 ± 1.5	111	1.51	0.11	16	5.75
3-B3	RA2	3.5	0.02	2.0	55	83	110*	0.75*	0.25	7	5.5 ± 1.5	111	1.51	0.11	16	3.5

3-B4	RA2	3.5	0.02	2.0	55	94	110*	0.85*	0.13	7	5.5 ± 1.5	111	1.51	0.11	16	3.75
3-B5	RA2	3.5	0.02	2.0	55	110	110*	1*	0	7	6.0	102	1.32	0.095	15	3.0
4-E1	BD	-	0.03	3.5	55	43.5	225*	0.19*	0.84	7	8±1	76	0.86	0.20	11	1.75
4-E2	BD	-	0.03	3.5	55	82	225*	0.36*	0.62	7	8±1	76	0.86	0.20	11	1.75
4-E3	BD	-	0.03	3.5	55	140	225*	0.62*	0.33	7	8±1	76	0.86	0.20	11	2.75
4-E4	BD	-	0.03	3.5	55	172	225*	0.76*	0.27	7	8±1	76	0.86	0.20	11	3.25
4-E5	BD	-	0.03	3.5	55	198	225*	0.88*	0.15	7	8±1	76	0.86	0.20	11	2.25
4-E6	BD	-	0.03	3.5	55	225	225*	1*	0	7	8±1	76	0.86	0.20	11	1.75
2-Ax	RA1	0.4	0.0115	0	24	0 to150	150+	1+	0	2	2.9	92	1.72	0.10	31	1.0
2-Bx	RA1	0.4	0.02	0	55	0 to350	350+	1+	0	7	5.0	122	1.74	0.087	18	0.75
2-Dx-a	RA1	0.4	0.003	0	55	1.7	10*	0.17*	0.80	2	7.0	87	1.05	0.064	13	15
2-Dx-b	RA1	0.4	0.003	0	55	4.1	10*	0.41*	0.57	2	7.0	87	1.05	0.064	13	15
2-Dx-c	RA1	0.4	0.003	0	55	6.7	10*	0.67*	0.30	2	7.0	87	1.05	0.064	13	15
2-Dx-d	RA1	0.4	0.003	0	55	9.1	10*	0.91*	0.17	2	7.0	87	1.05	0.064	13	15
2-Dx-e	RA1	0.4	0.003	0	55	10	10*	1*	0	2	7.0	87	1.05	0.064	13	15

For run names, the first number indicates bedrock type: 1-LG, 2-RA1, 3-RA2, and 4-Boulders. The letter indicates slope ranges and the last number indicates the sediment supply to capacity ratio in that slope value. S denotes slope, D denotes grain size, Q_w denotes water discharge, Q_s denotes sediment feed rate, H denotes mean water depth, U denotes mean flow velocity, $\boldsymbol{Fr}$ denotes Froude number $= U/\frac{Q_w}{\sqrt{gH,}}$ τ^* denotes the Shields number, B/H denotes the width-depth ratio, *tr* denotes the run time, Q_c denotes the capacity sediment transport rate, the ratio q_s/q_c denotes supply to capacity ratio, where *qs* = the volume sediment supply rate per unit width and q_c = the capacity sediment transport rate per unit width, and *po* denotes the fraction of the bed surface exposure. For the capacity sediment transport rate (Q_c) and the supply to capacity ratio(q_s/q_c), the annotation * indicates $Q_{al,c}$ and $q_s/q_{al,c,}$ respectively, while the annotation $^+$ indicates $Q_{a,c}$and $q_s/q_{a,c,}$ respectively.

[20] The Shields number τ* of Table 1 is defined as

$$\tau^* = \frac{HS}{RD} \tag{4}$$

where R denotes the submerged specific gravity of the sediment and D denotes mean grain size. The specific gravity of the sediment was assumed to be 2.65, so that R is assumed to take the value 1.65.

4.2. Capacity Transport Conditions

[21] To determine the capacity transport rate *Qal,c* from the experiments, we started any given run with some antecedent alluvium (except runs with S < 0.005), and gradually increased the sediment feed rate until the mixed alluvial-bedrock bed was completely covered by alluvium. In this way, we can make sure that we have an accurate and objective measurement of the capacity transport rate, rather than simply predicting it from a bedload relation such as *Meyer-Peter and Müller* [1948], a revised version given in *Wong and Parker* [2006].

[22] The flow depths were measured for all capacity flow conditions except for slope = 0.04 and 0.053 because of the extremely high dynamics of gravel movement and shallow supercritical flow. The capacity bedload transport rate per unit width *qal,c* can be characterized in terms of a dimensionless Einstein number;

$$q^* = \frac{q_{alc}}{\sqrt{RgDD}} \tag{5}$$

In Figure 7a, q^* is plotted against $\tau^* - \tau c^*$, where τc^* is critical Shields number. The data are compared with a) data from *Wong et al.* [2008], which used the same sediment as the 7 mm material used here, but in a narrower flume (channel width B = 0.50 m) in order to suppress alternate bars. In addition, the bedload transport relation obtained by *Wong et al.* [2008], i.e., is shown.

$$q^* = 2.66\,(\tau^* - \tau_c^*)^{1.5}$$

$$\tau_c^* = 0.0549 \tag{6}$$

[23] Figure 7a shows a reasonable correspondence between the capacity bedload transport data determined here, that of *Wong et al.* [2008] and equation (6). The points from the present study plot slightly scattered. This may be because of the fact that the width–depth ratio of the present capacity runs, which ranged from 11 to 25, was substantially larger than those of *Wong et al.* [2008], which ranged from 4 to 7.

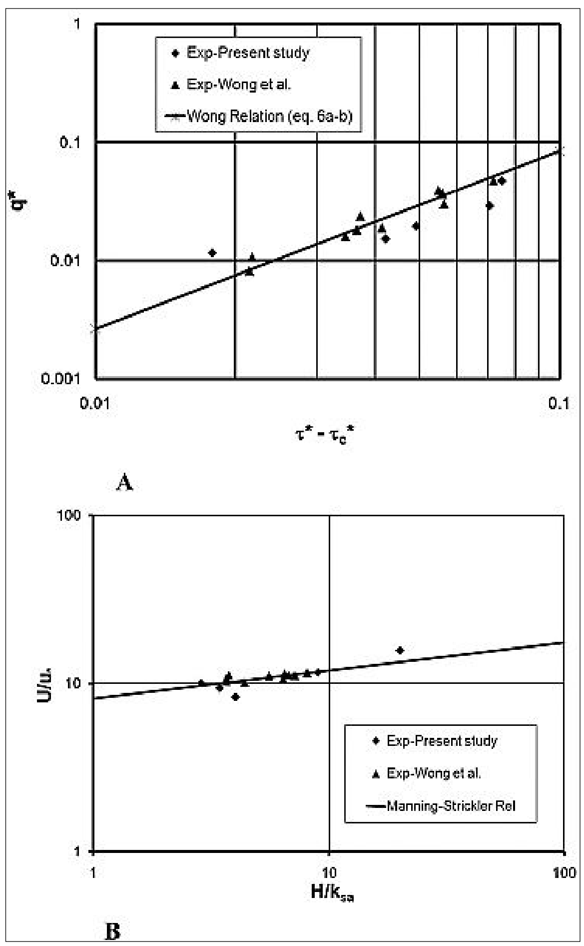

Figure 7.Comparison of the results from the present study and *Wong et al.* [2008] ,Phairot Chatanantavet ,and Gary Parker (2008) (a) capacity bedload transport rate in terms of a dimensionless Einstein number and (b) values of U/u_* versus H/ksa.

A consequence of this difference in width–depth ratio was that alternate bars were present in our experiments (and thus spatial variations in water depth and velocity), but absent in those of *Wong et al.*[2008].

[24] The alluvial roughness height ksa of the capacity runs (with a completely alluvial bed) was estimated with the relation [e.g., *Kamphuis*, 1974]

$$k_{sn} = 2D_{90} \tag{7}$$

where *D90* is a size such that 90 percent of the sediment is finer. For the essentially uniform sediment of the present study, *D90* was estimated as 2.0 mm and 8.7 mm for the finer and coarser sediment, respectively. Figure 7b shows a plot of U/u_* versus H/ksa, including (1) the capacity data obtained in the present study, (2) the data from *Wong et al.* [2008], and (3) Manning–Strickler relation from equation (2), (but with ksa replacing ksb therein). It is important to distinguish the definitions between ksb and ksa. While ksb is hydraulic roughness height of a bedrock surface, ksa is hydraulic roughness height of a fully covered alluvial bed.

[25] The consistency between the two data sets and with equation (2) is seen to be excellent. The present data shows more scatter than that of *Wong et al.* [2008]. This again may be because of the presence of alternate bars in our data and their absence in *Wong et al.* [2008].

4.3. Gradual versus Runaway Alluviation

[26] Figure 8 shows a series of aerial photographs which illustrate the time evolution of bedrock exposure for the case of Run 1-A2. For this run S = 0.0115, mean grain size D = 2 mm, supply to capacity ratio $qs/qal,c$ = 0.29, width to depth ratio = 20, the bed configuration was LG, and the total run duration tr = 5.25 hours. Note that although the run was started from random gravel patches, a bar pattern soon emerged. The bars took the form of a longitudinal strip of alluvial cover that curved alternately from one side of the channel to the other. The bar strip clearly propagates downstream while maintaining a constant bedrock exposure through time.

[27] Figure 8 documents a pattern of gradual alluviation. However the pattern of alluviation was not always gradual. For all runs with a bed slope $S \geq \sim 0.005$ (all slope thresholds presented here reflect the specific conditions of our experiments and may not be independent of other variables such as higher bed roughnesses or a wider range of hydraulic conditions) that were commenced from a bare bed, no alluviation was observed at any sediment supply rate qs (so that the fraction po of the bed that was open bedrock = 1), until an "oversaturated" capacity value qo,c was attained (at which point po suddenly dropped to 0; Figure 9a). At this point, runaway alluviation occurred, and the bed immediately underwent significant aggradation.

[28] Figure 9a shows results from three experiments to support these claims, all commencing from a bare bed. The figure documents the variation of bed exposure po as a function of qs/qc for these runs. For Runs 2-Ax (S = 0.0115) and 2-Bx (S = 0.02), the sediment supply rate qs was increased in steps from 0 to an oversaturated value qo,c, but the bed always remained bare of sediment except when runaway alluviation occurred at $qs = qo,c$. Runs 2Dx-a to 2Dx-e (S = 0.003), on the other hand, showed gradual alluviation, with the bed exposure po gradually reducing toward 0 as qs approached qal,c. It is of interest to note that the decrease of po with increasing $qs/qal,c$, following a relation that is approximately linear.

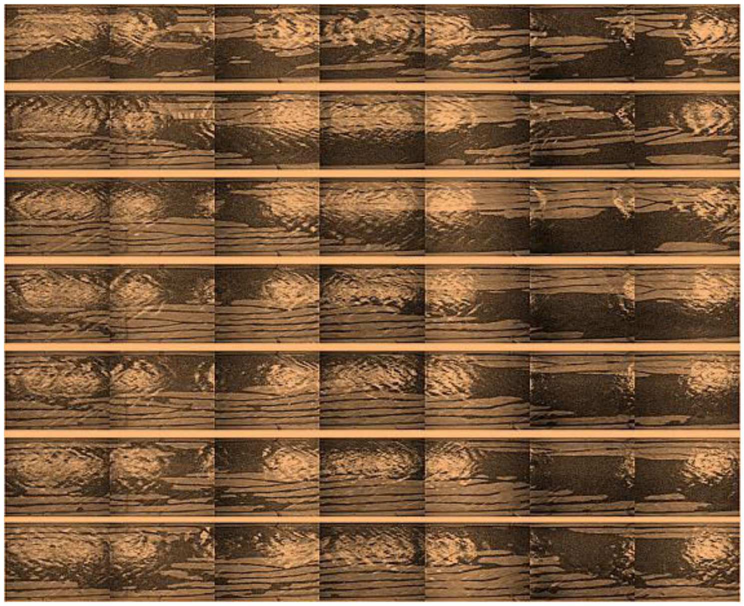

Figure 8.An example of the time evolution of bedrock exposure. The run in question is Run 1-A2; some relevant parameters are: S = 0.0115, $qs/qal,c$ = 0.29, and width/depth ratio = 20. The bedrock configuration is LG (longitudinal grooves). Times from top to bottom are 0.25, 1.25, 1.75, 2.25, 3.25, 4.25, and 5.25 hours. Flow from left to right. Note that the run was started from random hand-placed patches. Note also that sediment is dark and "bedrock" is light.

[29] In order to obtain partial bed cover, i.e., po < 1 for below-capacity conditions at these slopes greater than 0.015, it was necessary to begin with an antecedent condition such that the bed was completely covered with sediment to some depth hsi (as opposed to random patches of sediment in case of 0.005 < S< 0.015). If this initial depth of cover was below some critical value $hsic$, the antecedent alluvium would be completely washed off in time, and the run would continue to behave as if it had been commenced from a bare bed (Figure 9b). When hsi was above this value however the run would behave quite differently in four ways. Firstly, it became possible to attain equilibrium states with only partial coverage of the bed with sediment (po < 1). Secondly, the equilibrium value of fraction of the bed exposure po that remained bare bedrock decreased gradually and systematically with the sediment supply rate qs. Thirdly, the capacity sediment transport rate so obtained, here called qal,c, was significantly less than the value qo,c obtained from runaway alluviation of a bare bed. Fourthly, attainment of the capacity condition was not accompanied by bed aggradation, which could only be realized by further increasing the sediment supply rate qs.

[30] Figure 9b shows the time evolution of the fraction po of bed exposed for five runs, all with the same conditions (for example, S = 0.02, D = 7 mm, $qs/qal,c$ = 0.56) except for the antecedent thickness of sediment hsi. This value was equal to 0 cm for Run 2-B2-a, 1 cm for Run 2-B2-b, 2 cm for Run 2-B2-c, 4 cm for Run 2-B2-d, and 6 cm for Run 2-B2-e. In the case of no antecedent coverage, the bed remained bare throughout the run, in spite of continued sediment supply. In the case of hsi = 1 cm, all the sediment washed out in about 100 minutes, after which a bare bed was maintained in the presence of sediment supply. In the cases for which hsi = 2, 4 and 6 cm, in all cases the bed equilibrated to a value of po near 0.4. The implication is that the final state of a run is independent of the initial depth of alluvium hsi, as long as this value is above a critical value $hsic$.

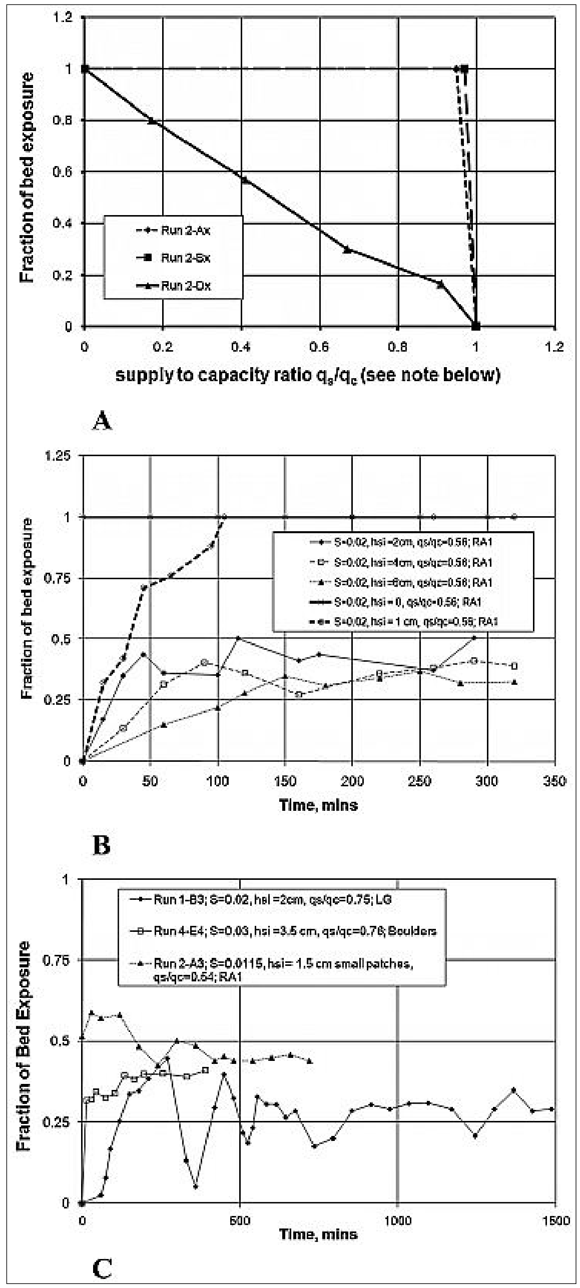

Figure 9. (a) Experimental results documenting the variation of bed exposure *po* as a function of *qs*/*qc*for runs commencing from a bare bed. Note that in the horizontal axis, *qs*/*qc* should be replaced by *qs*/*qo,c* in the case of runs 2-Ax and 2-Bx but replaced by *qs*/*qal,c* in the case of run 2-Dx. For Runs 2-Ax (*S* = 0.0115) and 2-Bx (*S* = 0.02), the sediment supply rate *qs* was increased in steps from 0 to *qo,c*, but the bed always remained bare of sediment except when *qs* = *qo,c*, when "runaway alluviation" [e.g. *Demeter et al.*, 2005] occurred. For Runs 2-Dx-a to 2-Dx-e (*S* = 0.003), the bed exposure changed gradually with *qs*/*qal,c*, following a relation that is approximately linear. (b) Experimental results showing the time evolution of bed exposure for six runs, Runs 2-B2-a to 2-B2-e, which had the same hydraulic and bedrock surface conditions but varying values of *hsi* corresponding to the initial depth of bed coverage with sediment. Note that for values of *hsi* ≤ 2 cm, the bed was eventually washed clean of sediment. For higher values of *hsi*, the bed eventually equilibrated to about the same fraction of bed exposure. (c) Experimental results showing the time evolution of bed exposure under different hydraulic and bedrock surface conditions. Results from Runs 1-B3, 4-E4, and 2-A3 are plotted.

[31] The characterization of $q_{o,c}$ as an oversaturated capacity transport rate should be clear from the explanation above; sudden alluviation was accompanied by sudden aggradation to a steeper slope. The precise value of the minimum thickness of antecedent sediment h_{sic} needed to obtain a gradual decrease in p_o with increasing q_s (rather than a bare bed progressing to runaway alluviation) proved to be an increasing function of bed slope S. It is important to recognize the essential non-uniqueness of the process of alluviation, i.e., a bare bed with runaway alluviation at $q_{o,c}$ versus gradual alluviation with increasing q_s, the difference solely depending on the initial condition of the run.

[32] For all runs with a slope $S < 0.015$, it always proved possible to commence the run with patches of sediment that did not completely cover the bed, and nevertheless obtain (a) values of $p_o < 1$ that gradually decreased with increasing q_s and (b) did not show runaway alluviation. At a slope S of 0.003, runaway alluviation never occurred, even when the run was commenced from a bare bed (Figure 9a). Instead, in this case gravel particles tend to collide with each other and form patches, a result that is similar to *Sklar and Dietrich* [2002] and *Demeter et al.* [2005].
[33] Figure 9c shows the time development toward equilibrium of three experiments: Run 1-B3 (S = 0.02, bed configuration = LG, D = 7 mm, h_{si} = 2 cm), Run 4-E4 (S = 0.03, bed configuration = RA1 with boulders, D = 7 mm, h_{si} = 3.5 cm), and Run 2-A3 (S = 0.0115, bed configuration = RA1, D = 2 mm, started with patches with average h_{si} = 1.5 cm). In all cases a reasonably constant mean value of p_o was attained after a sufficient amount of time. Increasing bed slope in the absence of boulders (for example, Run 2-A3 to Run 1-B3) appears to result in larger fluctuations about the mean at final equilibrium. The presence of boulders, on the other hand, seems to suppress fluctuations (Run 4-E4).

[34] If the antecedent cover of alluvium is sufficient, the degree of alluviation is a gradual function of sediment supply (will be shown in the sections 4.4 and 4.5) and the capacity sediment transport rate (q_c) equals the alluvial value ($q_{al,c}$). For $S > \sim 0.005$ however it is possible for the antecedent degree of alluvium to be insufficient for gradual alluviation. Instead the bed remains bare for all supply rates up to some oversaturated limit $q_{o,c}$, where in this case

$$q_c = q_{o,c} > q_{ol,c} \tag{8}$$

For $S < \sim 0.005$ however the process of alluviation is always gradual even with no antecedent cover of alluvium, where in this case

$$q_c = q_{ol,c} \tag{9}$$

Nonetheless, the slope threshold presented here reflects the specific conditions of our experiments and may not be independent of other variables such as higher bed roughnesses or a wider range of hydraulic conditions. However, as long as the hydraulic roughness of purely bedrock surface in question is lower than alluvial bed roughness of transported sediment (which is expected to be true most of the time in nature), the above slope threshold is expected to be applicable. Note that extreme bedrock channel topographies in nature such as inner channels do not necessarily mean higher bed roughness than the alluvial counterpart as a more systematic measure (for example, using the Manning–Strickler equation) needs to be considered.

[35] The behavior embodied in equations (8)–(9) suggests the following. At sufficiently low slopes, grain collisions are frequent, and they can result in the formation of equilibrium sediment patches regardless of the antecedent bed. At higher slopes however sediment rolling or saltating over a bare bed may either collide infrequently or energetically move, or in such a way that equilibrium deposits of sediment do not form.

4.4. Linear Relationship for Lower Slopes

[36] Sections 4.4–4.9 serve to summarize six main conclusions concerning the experiments. Note that in all the runs reported in these sections, antecedent gravel patches, or an antecedent gravel layer was emplaced, and a capacity transport rate *qal,c* was always eventually reached without runaway alluviation.

[37] Figure 10 shows the experimental results of bedrock exposure as a function of sediment ratio *qs*/*qal,c*for three bedrock configurations (LG, RA1, and RA2) with $S = 0.0115$ and $D = 2$ mm. While the data show some scatter according to bedrock configuration, the linear relation hypothesized by *Sklar and Dietrich* [1998, 2004] proves to be a reasonable rough approximation of the data. The results thus show that for lower slopes S (for example, $S < 0.015$) bedrock exposure *po* decreased more or less linearly with increasing *qs*/*qal,c*.

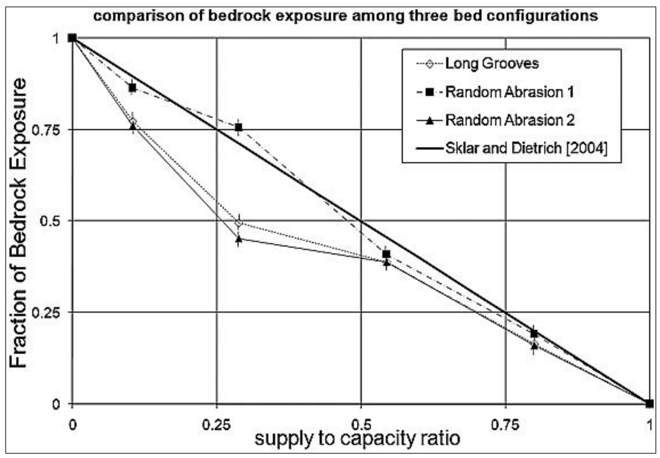

Figure 10. Experimental results showing bedrock exposure *po* as a function of the ratio *qs*/*qal,c* for various configurations of bedrock bed for $S = 0.0115$, and using 2-mm size gravel. The runs in question are Runs 1-A1 to 1-A5 (bed configuration LG), 2-A1 to 2-A5 (bed configuration RA1), and 3-A1 to 3-A5 (bed configuration RA2). All runs were commenced from an initial state with hand-placed patches. Also shown in the figure is the relation hypothesized by *Sklar and Dietrich* [1998, 2004].

[38] In the present experiments, the hydraulic roughness of the bedrock was less than that of a completely alluviated bed, or a partially alluviated bed covered with bars (and sometimes antidunes). As a result, for *qs*/*qal,c* < 0.5 it was found that bedrock roughness could play an important role in determining the degree of bedrock exposure. For 0.5 < *qs*/*qal,c* < 1.0, on the other hand, the composite roughness associated with grain roughness, bars, and/or antidunes dominates, so that the bedrock roughness no longer affects the degree of exposure of the bed.

[39] Figure 11 documents the process of gradual alluviation. The experiments in question are Run 1-A1, Run 1-A2, Run 1-A3, and Run 1-A4, for which the corresponding values of *qs*/*qal,c* were 0.10, 0.29, 0.54, and 0.80, respectively, and the corresponding values of *po* were 0.77, 0.50, 0.39, and 0.17, respectively. In all cases the experimental conditions were identical except for the sediment supply rate, with $S = 0.0115$, $D = 2$ mm, $Qw = 24$ l/s, and bedrock configuration LG.

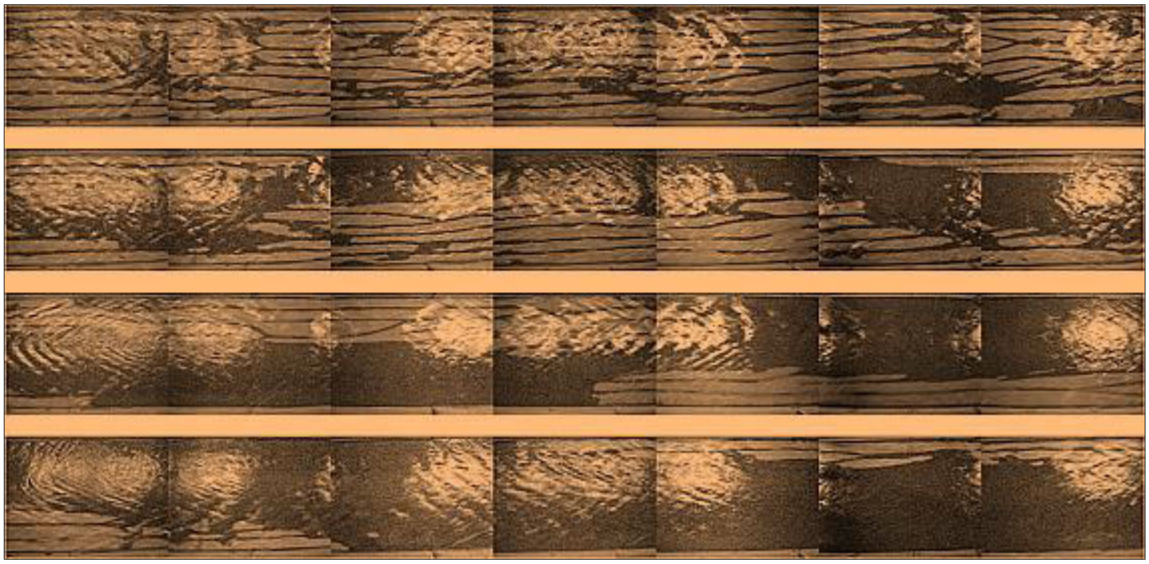

Figure 11.An example of the variation of equilibrium bedrock exposure with varied sediment supply when slope S = 0.0115, grain size D = 2 mm, water discharge Qw = 24 l/s, the bed configuration is LG, and $qs/qal,c$ takes the values = 0.10, 0.29, 0.54, and 0.80, from top to bottom, respectively. Note that each experiment with a given supply rate was run individually. The runs in question are Runs 1-A1 to 1-A4. The flow was from left to right. Note that sediment is dark and "bedrock" is light.

4.5. Full Bedrock Exposure versus Alluviation at Higher Slopes

[40] For sufficiently higher slopes (for example, S > 0.015) there is a range of values $qs/qal.c$ less than a critical value $(qs/qal.c)cr$ for which the bedrock remained fully exposed without any alluvial deposit regardless of the value of hsi. Within this range, the initial hand-placed gravel was all flushed away at sometime, and the particles saltated over the bed without coming to rest at equilibrium.

This range expands, i.e., $(qs/qal.c)cr$ increases with increasing slope (apparently regardless of shear stress, as noted below in section 4.6). However, within the upper range of values of $qs/qal,c$ (for example, where $qs/qal,c > (qs/qal,c)cr$), a linear relationship between the degree of exposure and $qs/qal,c$ still prevailed. Figure 12 illustrates bedrock exposure fraction po as a function of sediment ratio $qs/qal,c$ for all three bedrock configurations (LG, RA1, RA2) with S = 0.02 and D = 7 mm. Note the shift from full bed exposure at $qs/qal,c$ = 0.37 to a partially alluviated bed at $qs/qal,c$ = 0.56 for all three bedrock surface types.

[41] Experimental results showing bedrock exposure as a function of sediment ratio $qs/qal,c$ for various values of slope S are shown in Figure 13. The bed configuration is RA1. In the case S = 0.0115, po shows an essentially linear decrease with increasing $qs/qal,c$, with $(qs/qal,c)cr$ = 0. As the slope increases toward 0.053 however $(qs/qal,c)cr$ steadily increases. Once alluviation does occur however it tends to drop down to a linear relation that could be approximated by equation (1). Note that for S = 0.04 and 0.053, the Froude numbers are extremely high and the gravel movement is very energetic. It was too fast to measure many parameters aside from bedrock exposure, which is the main variable in question. For this reason, the Table 1 does not include the runs associated with these high slopes.

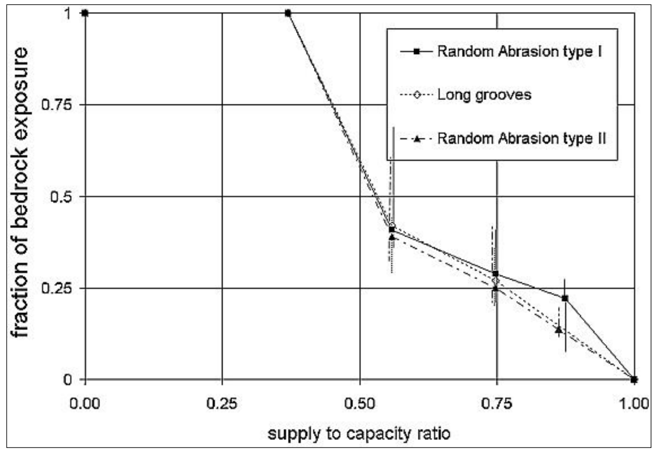

Figure 12. Experimental results showing bedrock exposure *po* as a function of sediment ratio *qs*/*qal,c* for all three bedrock types (LG, RA1 and RA2). The runs in question are Runs 1-B1 to 1-B4 (LG), Runs 2-B1 to 2-B5 (RA1) and Runs 3-B1 to 3-B5 (RA2). For all of the runs, S = 0.02, D = 7 mm gravel and the initial bed was completely covered with uniformly 2-cm thick gravel.

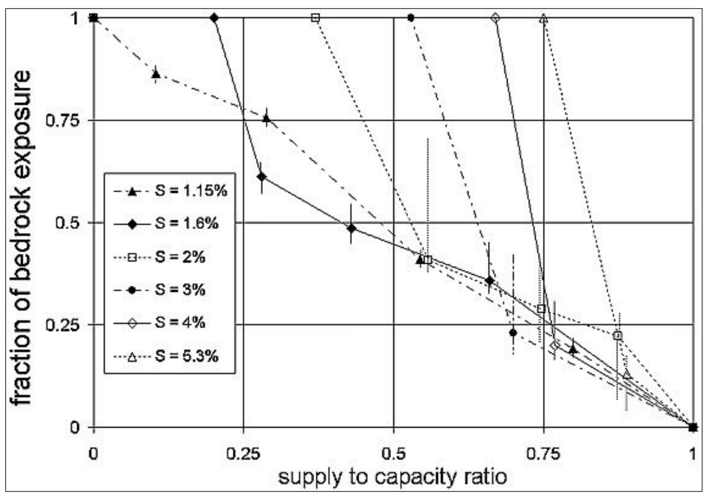

Figure 13. Experimental results showing bedrock exposure *po* as a function of the ratio *qs*/*qal,c* for various values of slope S using the bed configuration RA1. The slope S was equal to 0.0115 for Runs 2-A1 to 2-A5, 0.016 for Runs 2-C1 to 2-C4, 0.020 for Runs 2-B1 to 2-B5, and 0.030 for Runs 2-E1 to 2-E3.

4.6. Runs for Two Different Values of Dimensionless Shear Stress but Same Slope

[42] Two sets of experiments were compared for which bed slope S were both equal to 0.016, but the Shields number τ^* varied by a factor of more than two. Here we changed the Shields number by adjusting water discharge only. The experiments in question are Runs 3-C1 to 3-C5 (τ^* = 0.21) and 3-C6 to 3-C9 (τ^* = 0.09). In both cases the bed configuration was RA2. The variation of bedrock exposure fraction *po* with *qs*/*qal.c* shown in Figure 14 is found to be similar for both runs. The results suggest that *po* is a strong function of *qs*/*qal,c* and S, but perhaps a much weaker function of Shields number.

[43] Having noted the similarity in Figure 14 however there was a difference between the two runs. In the run with a higher value of τ^*, antidunes tended to dominate over bars, whereas in the run with the lower value of τ^* bars dominated over antidunes.

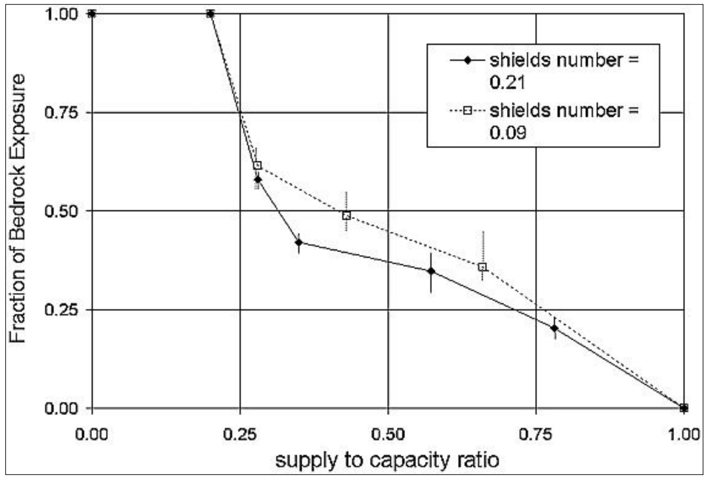

Figure 14.Comparison of two sets of experimental runs with different dimensionless shear stresses τ^*, but constant slope (0.016) using the bed configuration RA2. The runs in question are Runs 3-C1 to 3-C5 (τ^* = 0.21) and 3-C6 to 3-C9 (τ^* = 0.09). All runs were commenced with complete coverage of the bed with sediment.

4.7. Influence of the Addition of Model Boulders

[44] For sufficiently high slopes (for example, $S > 0.02$), the addition of model boulders (which considerably increase hydraulic roughness) in the channel (Figure 4a) suppresses the over-exposure of the bed at the lower range of values of $qs/qal,c$ associated with saltating grains that fail to form deposits. The boulders were seen to cause deposition in the form of alluviated zones due to the formation of multiple local hydraulic jumps (Figure 4b). As a result, a linear relation between the degree of exposure po and sediment ratio $qs/qal,c$ was restored by the addition of boulders. Figure 15 shows the experimental results for bedrock exposure fraction po as a function of the ratio $qs/qal,c$ for runs without and with boulders, at a slope S of 0.03. The value of $(qs/qal,c)cr$ is greater than zero for the case without boulders, but equal to zero for the case with boulders.

[45] It can be seen from Figure 15 that the bed exposure fraction po is less than 1 even when the sediment supply rate is vanishing. This simply reflects the fact that the boulders covered a fraction 0.11 of the bed area.

[46] Figure 16 shows top views of the degree of alluvial bed cover realized for five experiments with boulders. The following parameters apply to the experiments: $S = 0.03$, and the ratio $qs/qal,c$ = 0.19, 0.36, 0.62, 0.76, and 0.88, from top to bottom, respectively. The runs in question are Runs 4-E1 to 4E-5. The gradual process of alluviation is readily apparent from the figure. Note the similarity in the gravel bar morphology induced by boulders between the field, as illustrated in Figure 5, and the experiments shown in Figure 16.

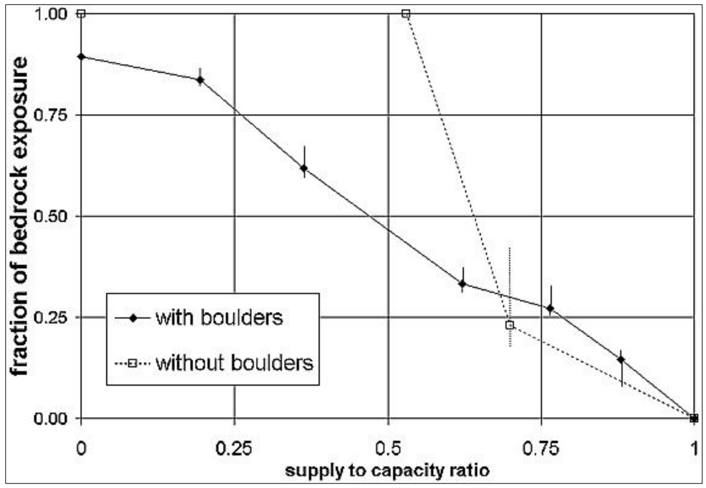

Figure 15. Experimental results showing bedrock exposure *po* as a function of the ratio *qs*/*qal,c* for runs without boulders and with boulders for slope = 0.03. The runs without boulders correspond to Runs 2-E1 to 2-E3, and the runs with boulders correspond to Runs 4-E1 to 4-E6. The bedrock surface configuration was RA1. All runs were commenced with complete cover. The fact that *po* is not equal to 1 for the case of *qs*/*qal,c* = 0 with boulders reflects the fact that the boulders themselves occupied a fraction of 0.11 of the bed surface.

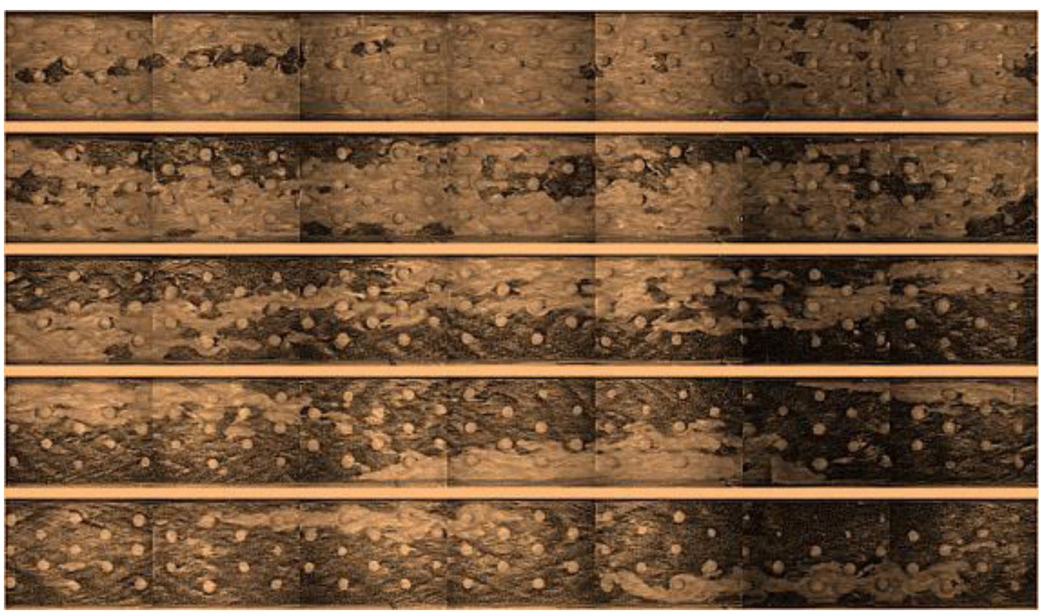

Figure 16. An example of equilibrium bedrock exposure with varying sediment supply in the presence of boulders. The following parameters apply to the experiments: *S* = 0.03 and the ratios *qs*/*qal,c* = 0.19, 0.36, 0.62, 0.76, and 0.88, from top to bottom, respectively. Note that each experiment with a given supply rate was run individually. The runs in question are Runs 4-E1 to 4-E5. The flow was from left to right.

4.8. Depth of Alluvial Deposit

[47] The degree of alluviation of the bed can be characterized not only in terms of the fraction of the bed *po* that is free of alluvium, but also in terms of the average depth *ha* of alluvium. Here this depth is averaged only over the zone where alluvium is present, i.e., it does not include zones where the bed is bare. Over an appropriate range of sediment supply conditions,

the value of *ha* associated with mobile-bed equilibrium increases linearly with *qs/qal,c*. This is illustrated in Figure 17. This behavior is documented for both below-capacity alluvial deposition on a smooth bedrock bed, as well as on a bed with model boulders.

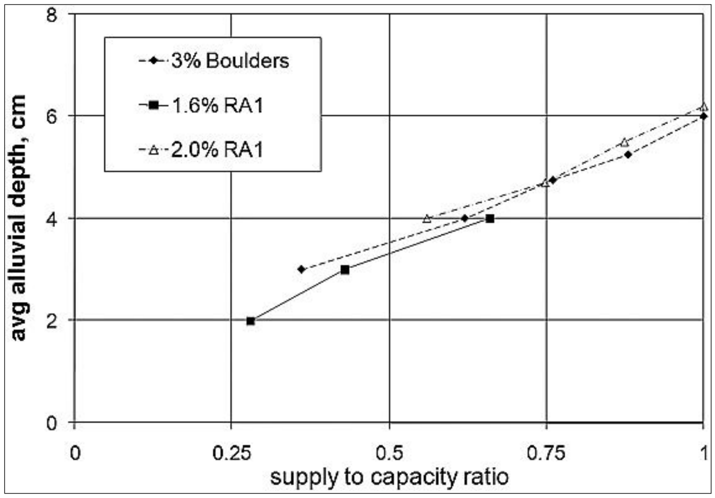

Figure 17.Experimental results showing a nearly linear relationship between the average gravel depth at equilibrium and the ratio *qs/qal,c*. Here the average gravel depth is computed over that part of the bed actually covered with gravel, not the entire bed. The runs in question are Runs 2-C1 to 2-C3 (S = 0.016, bed configuration RA1), Runs 2-B2 to 2-B5 (S = 0.02, bed configuration RA1), and Runs 4-E2 to 4-E6 (S = 0.03, bed configuration RA1 with boulders).

4.9. Formulations for Use in Landscape Evolution Models

[48] On the basis of the results in the previous sections, we can tentatively propose two types of formulations for estimating the degree of bedrock exposure *po* as a function of sediment supply ratio for use in landscape evolution models.

[49] A fairly simple model can be delineated as follows. For bedrock streams where large boulders (or perhaps extreme topographies that are much rougher than the alluvial bed based on the transported grain size) are present, it is reasonable to use a linear relation between the fraction of bedrock exposure *po* and the sediment supply ratio *qs/qal,c* [e.g., *Sklar and Dietrich*, 2004], i.e., equation (1). The estimation of capacity rate *qal,c* in the case of streams with large boulders can be seen in *Yager et al.* [2007]. For bedrock streams where large boulders (or extreme topographies) are not present and channel slopes are high (for example, $S > \sim 0.015$), such as streams in some badland settings, it is fairly reasonable to use a relation with an abrupt shift as follows:

$$p_0 = \begin{cases} 1, & q_s < q_{al,c} \\ 0, & q_s = q_{slc} \end{cases} \tag{10}$$

In the case for which boulders are not present however the more advanced tentative relation can be developed based on the results of Figure 13:

$$p_0 = \begin{cases} 1, \dfrac{q_s}{q_{al,c}} \leq \left(\dfrac{q_s}{q_{al,c}}\right)_{cr} \\ 1 - \dfrac{q_s}{q_{al,c}}, \dfrac{q_s}{q_{al,c}} > \left(\dfrac{q_s}{q_{al,c}}\right)_{cr} \end{cases} \tag{11}$$

The value of (*qs*/*qal,c*)*cr* appears to increase solely with channel slope and can be obtained from Figure 13. These values are reproduced in Figure 18. Note again that all the experimental runs used in that figure were started from antecedent hand-placed gravel (with *hsi* higher than *hsic*) uniformly distributed along the channel.

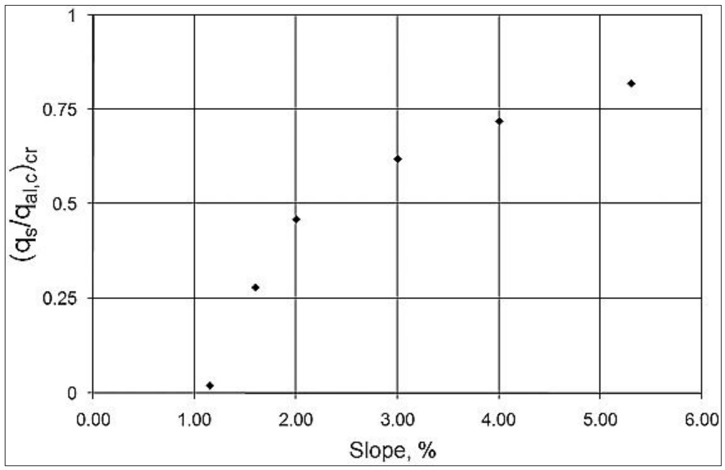

Figure 18. Experimental results showing the relationship between critical sediment supply to capacity ratio (*qs*/*qal,c*)*cr* and channel slope using the bed configuration RA1. The results are deduced from Figure 13.

[50] In implementing the above formulation, it is necessary to have a relation to predict *qal,c*. At least for the present experiments, equation (6) does an adequate job in this regard in the case without boulders.

5. Discussion

[51] The present study pertains to gravel of uniform size. The experiments reported here show that the degree of bedrock exposure depends primarily on channel hydraulic roughness, the sediment supply, and channel slope. The findings indicate that the relation governing the degree of bedrock exposure as a function of sediment supply should show a difference between channels of high slope and channels of low slope. The results illustrated in the section 4.5 thus offer a possible explanation as to why the bedrock bed is fully exposed without any alluvial coverage in some bedrock streams, such as the Waipaoa River, New Zealand [e.g., *Crosby and Whipple*, 2005].

[52] However, in a stream that transports finer gravel between and around large, immobile, colluvially derived boulders, these boulders can trap smaller particles among them, generate drag forces which reduce the transport capacity for the finer gravel and dissipate the energy of the flow through turbulence around the boulders. The results illustrated in section 4.7 thus suggest that in a setting where large boulders are present, a linearly decreasing relationship between bedrock exposure and sediment supply prevails, and might be valid

regardless of channel slope. This suggests the validity of the simple linear formula used in *Sklar and Dietrich* [1998, 2004].

[53] In an environment such as a badland landscape for which large boulders (or perhaps extreme topographies) are not present however the results illustrated in the Figures 9a and 9b, so called runaway alluviation combined with the results in section 4.5 support an abrupt shift from a fully exposed bedrock to a fully alluviated state when the sediment supply exceeds transport capacity. This concept is used to quantify the cover effect in the badland landscape evolution model of *Howard* [1994]. In the channels in a badland landscape observed by *Howard and Kerby* [1983], an abrupt downstream transition between purely bedrock and purely alluvial channels prevails. Figure 4 and Table 1 therein indicate a slope range 0.01–0.10 for alluvial channels and a corresponding range 0.015–0.60 for bedrock channels. Part of the present work may thus apply to their setting as well.

[54] The behavior in Figure 9a may have field analogies. In geologically young, uplifting mountain areas, sediment supply from hillslopes and tributaries may be sufficient to ensure that antecedent "sediment patches" always form between floods. This may prevent the occurrence of a bare bed followed by runaway alluviation as *qs* increases. In bedrock streams with extremely low sediment supply and low denudation rate of rocky hillslopes, such as Piccaninny Creek, Australia; [Figure 2 from *Wohl*, 1993] however the bed may remain bare even as sediment supply changes.

[55] Numerical modeling studies [e.g., *Crosby et al.*, 2007; *Gasparini et al.*, 2007] have used incision equations, based on the work by *Sklar and Dietrich* [2004], with tools and cover type models that include a linear relation between alluvial cover and *qs*/*qal,c*. These studies have shown a parabolic relationship between bed erodibility and *qs*/*qal,c*. Our results in sections 4.5 and 4.9 suggest that for steep channels with no big boulders or possibly extreme bed topographies, higher erodibilities at lower cover values (even with higher *qs*/*qal,c*) may result and potentially modify the parabolic relation that these studies have found. This discrepancy could potentially result in higher incision rates for steep landscapes than those of which these numerical models predicted.

[56] The result that a critical initial sediment thickness is required to avoid complete clearing of alluvium for a given sediment supply could be interpreted as follows. This process can be viewed as the opposite of runaway alluviation but it is instead runaway transport and bed exposure. In the runaway alluviation case, there is a positive feedback between increasing bed hydraulic roughness and more deposition, which decreases the transport capacity and leads to further deposition. The results shown in Figure 17 also suggest that the increase in average alluvial depth with increasing values of *qs*/*qal,c* is an example that bed roughness tends to increase with alluviation beyond just the sediment diameter itself. Since some bedrock patches are present (but not averaged over), the overall hydraulic bed roughness must increase with *qs*/*qal,c*.

[57] In the runaway bed exposure case, the positive feedback is that more bed exposure decreases the bed hydraulic roughness, increasing transport capacity and leading to more sediment being entrained from the bed until it is bare. These behaviors depend on bare bedrock being smoother than an alluviated bed (with the bed particle size of the transported sediment). Our study thus is limited to this condition, in which we believe it is more common in nature.

[58] The results in section 4.3 also pose a question: "Why doesn't the initial sediment wash out regardless of the initial sediment thickness?" The answer might be alluvial bed

roughness (in the form of bars or antidunes). Thicker layers of alluvium perhaps lead to higher alluvial bedform amplitudes. This additional roughness that develops from thicker initial sediment layers sufficiently reduces the transport capacity of the flow to balance the imposed incoming sediment flux *qs*, and runaway exposure does not occur. This hypothesis illustrates how bed roughness can be a degree of freedom for channels to adjust to imposed sediment flux conditions.

[59] The present study is only limited to relatively planar bed configurations. In other words, our experiments do not include extreme topographies (for example, irregularly inner channels), which may or may not have higher hydraulic roughness than the alluvial counterpart. Our experiments are also limited to non-erodible beds only in order to keep the bedrock bed roughness constant throughout the study.

[60] From the results, we hypothesize here that bed roughness in Bedrock Rivers (*ks*) (for example, from bedrock surface, bedforms such as bars or step-pool, alluvium, coarse alluvium such as immobile boulders) is a first-order control on the patterns of deposition, sediment entrainment, and thus bedrock exposure. In other words, the hydraulic roughness in a bedrock channel has a positive feedback with sediment deposition. Primary dimensionless numbers involving in this study should include *Fr*, *ks*/*D*, and *S*. These could be verified systematically in future study.

6. Conclusion

[61] The present experimental work consists of a detailed investigation of the factors controlling the degree of bedrock exposure in mountain streams. In particular, it quantifies how bedrock exposure varies with sediment supply. The results suggest that the sediment supply, channel slope, hydraulic bed roughness, the intensity/thickness of antecedent gravel in the channel, and the presence of boulders (which can be viewed as a form of bed roughness) in the channel are major factors controlling whether a bedrock surface is fully exposed or partially alluviated.

[62] The present study indicates that the formulations of the cover effect used in previous landscape evolution models [e.g., *Sklar and Dietrich*, 1998, 2004; *Howard*, 1994; *Tucker and Slingerland*, 1994] are reasonably valid within appropriate but different ranges. To be specific, the simple linearly decreasing relation between the degree of bedrock exposure and the ratio of sediment supply rate to capacity transport rate used in the saltation–abrasion model of *Sklar and Dietrich* [1998, 2004] was found to apply when the channel contains colluvial boulders (i.e., dramatically high bed roughness), and also when the bed slope is very low (less than 0.005 for our experimental setting).

[63] In contrast, the concept of an abrupt shift from a fully exposed bedrock channel to a fully alluviated channel when the sediment supply exceeds the transport capacity, as used in the landscape evolution models of *Howard* [1994] and *Tucker and Slingerland* [1994], also seems to have a range of validity, corresponding to sufficiently high slopes (S > 0.015 for our experimental setting) and the absence of non-mobile boulders.

[64] The results of the analysis include (1) a relation for uniform sediment that can be used to compute the alluvial (as opposed to oversaturated) capacity gravel transport rate and (2) two relations for predicting the degree of exposure of the bedrock bed as a function of sediment supply and slope. The present analysis must be considered tentative because only two uniform grain sizes have been considered. It does however offer tools for improving models of landscape

evolution in which incision into bedrock plays an important role. In addition, it suggests useful avenues for further improving our understanding of the morph dynamics of below-capacity gravel transport in bedrock streams.

Notation

B channel width

Dcharacteristic grain size

D_{90} grain size such that 90 percent of the sediment is finer

F_r Froude number

g gravitational acceleration

H mean water depth

h_a mean depth of alluvial cover, computed with the exclusion of bare part of bed

h_{si} mean depth of initial hand-placed gravel patches

h_{sic} critical value of *hsi* below which an equilibrium alluvial cover cannot be maintained

k_s hydraulic bed roughness in bedrock rivers (for example, from bedrock surface, bedforms such as bars or step-pool, alluvium, coarse alluvium such as immobile boulders)

k_{sa} hydraulic roughness of the grains of a completely alluviated surface

k_{sb} hydraulic roughness of the bedrock surface

p_0 areal fraction of bedrock exposure

Q_c sediment transport capacity rate

$Q_{al,c}$ sediment transport capacity rate approached via gradual alluviation of the bed as the sediment supply increases

$Q_{o,c}$ oversaturated sediment transport capacity rate for the case of sudden alluviation from a sediment-free bedrock surface

q_c sediment transport capacity rate per unit width

$q_{al,c}$ sediment transport capacity rate per unit width approached via gradual alluviation of the bed as the sediment supply increases

$q_{o.c}$ Oversaturated sediment transport capacity rate per unit width for the case of sudden alluviation from a sediment-free bedrock surface

Q_s sediment supply rate

q_s sediment supply rate per unit width

$(q_s/q_{al,c})_{cr}$ critical sediment supply to capacity ratio at which the bedrock exposure changes abruptly from full bed exposure to partial alluviation

Q_w water discharge

q^*dimensionless Einstein bedload transport rate, as defined in equation (5)

R submerged specific gravity of sediment (equal to 1.65 for quartz)

Schannel slope

t_r run time of an experiment

U mean flow velocity

$u*$ shear velocity defined in equation (3)

τ* Dimensionless Shields number defined in equation (4)

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