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Flow resistance, sediment transport, and bedform development in a steep gravel-bedded river flume

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ABSTRACT

Quantifying flow resistance and sediment transport rates in steep streams is important for flood and debris flow prediction, habitat restoration, and predicting how mountainous landscapes evolve. However, most studies have focused on low gradient rivers and the application of this work is uncertain for steep mountain streams where surface flows are shallow and rough, subsurface flows are not negligible, and there is form-drag from bed- and channel-forms that differs from those in low gradient rivers. To evaluate flow resistance relations and sediment transport rates for steep channel beds, experiments were conducted using a range of water discharges and sediment transport rates in a 12 m long recirculating flume with bed slopes of 10%, 20%, and 30%, and a bed of nearly uniform natural gravel. Flow resistance for planar beds and beds that developed bedforms match empirical models that account for bedload-dependent roughness. Some bedforms were atypical for natural rivers at these bed slopes, such as stepped alternate bars and upstream migrating alternate bars. Total flow resistance increased with decreasing particle submergence and energetic sediment transport and drag on bedforms. Using linear stress partitioning to calculate bed stresses due to grain resistance alone, sediment flux relations developed for lower gradient rivers perform well overall, but they overestimate fluxes at 20% and 30% gradients. Based on previous theory, mass failure of the bed, which did not occur, was predicted for the highest Shields stresses investigated at 20% and 30% bed slopes; instead a concentrated layer, four to ten particle diameters deep, of highly concentrated granular sheetflow was observed.

1. Introduction

Steep mountain streams are an important component of the river network, as they provide aquatic habitat (Church, 2002), are conduits for sediment delivered to lower gradient channels (Milliman and Syvitski, 1992; Yager et al., 2012), and they comprise much of the channel network in mountainous regions (Shreve, 1969; Stock and Dietrich, 2003). Understanding steep stream hydraulics and sediment transport is therefore important for flood and debris flow prediction and mitigation, channel engineering and restoration, and landscape evolution (Buffington et al., 2004; Jakob et al., 2005; Takahashi, 2007; Rickenmann and Recking, 2011). However, most studies on river processes have focused on lower gradient rivers and flume studies ($S < 1\%$, where S is the tangent of the bed slope angle, θ), leaving uncertainty about whether relations developed for flow resistance and bedload fluxes can be applied to steeper channels (Scheingross et al., 2013; Schneider et al., 2015). For example, it has been suggested that flow resistance coefficients (C_f), which relate bed shear velocity ($u^* = \sqrt{\tau_b/\rho}$, where τ_b is the bed stress and ρ is the density of water) to the depth-averaged water flow velocity (U) (i.e., $C_f = u^{*2}/U^2$), in steep streams are much greater than empirical models predict (Bathurst, 1985; Wilcox et al., 2006; Ferguson, 2007; Rickenmann and Recking, 2011). This has been hypothesized to be due to increased form drag caused by pressure differentials around immobile clusters of grains, boulders, bedforms or woody debris (Wilcox et al., 2006; Yager et al., 2007; Ferguson, 2012). Alternatively, it has been shown that flow resistance increases in shallow, rough flows due to changes in the velocity profile near a rough bed (Lamb et al., 2017a, 2017b). Similarly, sediment transport rates are thought to be different in steep streams, as the presence of immobile (or rarely mobile) boulders, particle clusters, and channel forms, such as step-pool sequences, may stabilize sediment (Church et al., 1998; Chin and Wohl, 2005; Yager et al., 2007; Zimmermann et al., 2010; Prancevic and Lamb, 2015). In addition, increased form drag may reduce available shear stress for entraining and transporting sediment, leading models to under-predict critical shear stresses for initial motion of the bed (τ_c^*) and over-predict sediment fluxes (Rickenmann, 1997; Yager et al., 2007; Mueller et al., 2008; Nitsche et al., 2011; Schneider et al., 2015). Lower lift coefficients and reduced turbulent intensities in steep, shallow flows may also cause reduced sediment transport rates (Lamb et al., 2008, 2017a, 2017b).

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Although bed- and channel-forms affect flow resistance and sediment transport (Hassan and Reid, 1990; Aberle and Smart, 2003; Nitsche et al., 2011), bedform stability regimes are still largely unknown for mountain streams (Wohl and Merritt, 2005; Zimmermann et al., 2010; Buffington and Montgomery, 2013; Palucis and Lamb, 2017). Field observations suggest that different channel morphologies can be attributed to distinct ranges in bed slope (Buffington and Montgomery, 2013). For example, Montgomery and Buffington (1997) showed that channel state changes from alternate bars to plane bed (or the absence of channel- or bed-forms) to steps and pools to cascade morphology with increasing channel bed slope. This is in contrast, however, to more mechanistic theoretical and experimental investigations into the formation of specific channel states, where variables such as the channel width (W) to flow depth (H) ratio or the Froude number ($Fr = U/(gH)^{0.5}$, where g is the acceleration due to gravity), not bed slope, are shown to control channel state (Colombini et al., 1987; Grant et al., 1990; Montgomery et al., 2003; Church and Zimmermann, 2007). Combining field data and theory, Palucis and Lamb (2017) showed that these controlling variables co-vary systematically with bed slope, but predicting which state will emerge under a given set of conditions is still unclear.

The difficulty in observing active sediment transport in steep streams, combined with the lack of data on steep river hydrodynamics under a wide range of flow conditions, has led some to conduct flume experiments aimed at measuring flow resistance and sediment fluxes at steep bed slopes ($1\% < S \leq 20\%$) (Smart and Jäggi, 1983; Bathurst et al., 1984; Cao, 1985; Graf et al., 1987; Rickenmann, 1990; Recking, 2010). Cao (1985) collected hydraulic data for slopes ranging from 1 to 9% and relative submergence (defined as the ratio of the flow depth to the bed roughness height, k_s, which often scales with grain diameter, D) between 1.3 and 14, and showed that the resistance coefficient increased with decreasing H/D. Smart and Jäggi (1983) and Rickenmann (1990) produced data for steeper bed slopes (up to $S = 20\%$) and a relative submergence of ~4, and saw increases in flow resistance with increasing sediment transport. Recking (2006) collected data in the same slope range as Cao (1985), but at higher bed shear stresses, in order to isolate the effect of bedload transport on C_f at a given H/D. In most previous work, the sediment bed was maintained at planar or near-planar conditions (i.e., no bedforms) and the flow depth was often deeper than a sediment diameter (i.e., relative submergence > 1). Mizuyama (1977) and Bathurst et al. (1984) conducted some of the few flume experiments on 20% bed slopes, with bed stresses high enough to develop bedforms, namely anti-dunes and alternating bars, and relative submergence as low as ~0.7. Mizuyama (1977) did not observe differences in flow resistance coefficients between a plane bed and one with alternating bars, but he did see slight increases in C_f with the onset of thalweg sinuosity. He also found that flow velocity did not change significantly with the onset of sediment transport. Lamb et al. (2017a) explored the effect of steep bed slopes (up to $S = 30\%$) and shallow flows (submergence down to 0.1) on flow resistance over a fixed, planar bed, but in the absence of sediment transport. They found that flow resistance matched observations in natural steep streams, despite the lack of bed- or channel-forms, suggesting that grain drag can account for much of the observed flow resistance in steep natural streams.

Another complicating factor in understanding very steep rivers is whether fluvial processes (where fluid-particle interactions result in rolling, saltation, or dilute suspensions of grains (Shields, 1936)), or mass flow processes dominate in the range of $10\% < S < 30\%$ in natural channels. The model proposed by Takahashi (1978) for in-channel bed failure assumes that when applied shear stresses (due to parallel seepage and surface flow) overcome resisting stresses within a granular bed at some depth, δ, particles above δ move together (*en masse*). Prancevic et al. (2014) showed through flume experiments that there exists a critical slope (S_c), defined as the slope above which in-channel failures occur prior to any bedload transport. They suggest that mass failure of channel beds might occur at slopes lower than the critical slope if the dimensionless bed stress, or the Shields stress ($\tau^* = \frac{\tau_b}{(\rho_s - \rho)gD}$, where ρ_s is the sediment density), is substantially higher than the critical value for fluvial transport (e.g., Shields stresses approaching one), based on the model of Takahashi (1978). While very few studies have investigated this regime, Smart and Jäggi (1983) observed that for $S = 20\%$, the mode of transport transitioned at high Shields stresses ($\tau^* \sim 0.69$) such that it became difficult to distinguish between bedload and suspended load. Mizuyama (1977) also found that for $S = 20\%$ and $\tau^* \to 1$, the mode of transport changed, such that the upper portion of the bed began to 'creep', which he referred to as an 'immature' debris flow, and others have described as a debris flood (Hungr et al., 2014). While these studies suggest that a transport transition may occur at high τ^*, the observations and data are sparse, especially for $S > 20\%$.

There is a need to acquire flow resistance and sediment flux data at steep slopes ($S \geq 10\%$) under high bed stresses in order to test whether fluvial or debris flow transport processes dominate (sensu Prancevic et al., 2014), and to determine if commonly used flow resistance models (Ferguson, 2007; Recking et al., 2008) and sediment flux models (Recking, 2010; Schneider et al., 2015) are broadly applicable to very steep channels. To address this need, a series of steep flume experiments were conducted in the Earth Surface Dynamics Laboratory at the California Institute of Technology. A subset of data from these experiments appears in a companion paper, Palucis et al. (2018), which focused specifically on the development of sheetflow at high Shields numbers, the structure of particle velocities within sheetflows, and how these flows differ from sheetflow occurring in low gradient systems or dry granular flow. In this contribution we present new data on the development of bedforms, flow resistance and sediment transport across the bedload-to-sheetflow transition. Our major objectives were to (1) determine whether debris flows could initiate via mass failure of the bed under uniform water flow conditions, (2) characterize the evolution and morphology of bedforms on steep slopes, (3) determine how these bedforms affect flow resistance and sediment fluxes, and (4) test flow resistance and sediment flux relations. The methods and experimental setup are discussed in Section 2, and the flow resistance and sediment flux data, as well as a detailed characterization of the bed state under each equilibrium flow condition, are presented in Section 3. Section 4 discusses how bed states differ from low-sloping rivers and affect flow resistance and sediment fluxes, and Section 5 is a summary of our findings.

2. Experimental setup and methods

As testing the bed failure model of Takahashi (1978) and Prancevic et al. (2014) was a major goal, a large flume width-to-grain diameter ratio ($W_{fl}/D_{84} = 29.5$, where D_{84} is the grain size for which 84% of the grains are smaller) was chosen to suppress the development of granular force chains that might cause grain jamming with the side walls, which could inhibit bed failure (Jop et al., 2005; Prancevic et al., 2018). This condition also likely suppressed the formation of step pools (Church and Zimmermann, 2007). A thick sediment bed, relative to the flow depth, was chosen to allow for a wide range of possible bed failure-plane depths (Takahashi, 1978; Prancevic et al., 2014). This experimental setup is unlike some natural mountain stream beds that have a thin veneer of large boulders over a relatively narrow bedrock channel, pieces of coarse woody debris, and other roughness elements (Montgomery and Buffington, 1997). These attributes were deliberately not included in our experiments to focus on conditions for bed failure in order to isolate the effect of channel bed slope and Shields stress on bedform development and sediment fluxes in a simplified system. The experimental setup is more directly analogous to mountain channels in arid landscapes, especially those that have recently experienced a large input of sand or fine gravel from landsliding and bank failures (Coe et al., 2008; Berger et al., 2011; McCoy et al., 2012), or following