Incipient sediment motion across the river to debris-flow transition

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ABSTRACT

Sediment transport in mountain channels controls the evolution of mountainous terrain in response to climate and tectonics and presents major hazards to life and infrastructure worldwide. Despite its importance, we lack data on when sediment moves in steep channels and whether movement occurs by rivers or debris flows. We address this knowledge gap using laboratory experiments on initial sediment motion that cross the river to debris-flow sediment-transport transition. Results show that initial sediment motion by river processes requires heightened dimensionless bed shear stress (or critical Shields stress) with increasing channel-bed slope by as much as fivefold the conventional criterion established for lowland rivers. Beyond a threshold slope of ~22°, the channel bed fails, initiating a debris flow prior to any fluvial transport, and the critical Shields stress within the debris-flow regime decreases with increasing channel-bed slope. Combining theories for both fluvial and debris-flow incipient transport results in a new phase space for sediment stability, with implications for predicting fluvial sediment transport rates, mitigating debris-flow hazards, and modeling channel form and landscape evolution.

INTRODUCTION

Outside of glaciated regions, channel morphology (e.g., Montgomery and Buffington, 1997) and landscape response to changes in climate and tectonics (e.g., Howard, 1994; Stock and Dietrich, 2003) are determined by sediment transport within channels by rivers and debris flows. Fluvial sediment transport occurs through fluid-particle interactions in rivers that result in rolling, saltation, or dilute suspensions (e.g., Shields, 1936). Debris flows, on the other hand, are highly concentrated slurries where solid and fluids are intermixed and both influence motion (e.g., Iverson et al., 1997). Although the physics of fluvial and debris-flow transport are distinct, we lack observations of when sediment moves in very steep channels and where initial sediment motion by one mode of transport dominates over the other. Consequently, most landscape-scale models do not differentiate these two important processes (e.g., Howard et al., 1994), and debris-flow hazard predictions rely on site-specific, multiple-regression techniques (e.g., Coe et al., 2008).

There is a paucity of data on sediment motion in channels steeper than $\theta = 6°$, where θ is the channel-bed angle. Classic theoretical models for initial sediment motion by river processes indicate that sediment transport occurs at lower near-bed fluid stresses (τ) with increasing bed slope due to the increased component of gravity acting on sediment in the downstream direction (e.g., Wiberg and Smith, 1987), consistent with experiments in sealed ducts (e.g., Chiew and Parker, 1995). Limited field and experimental data (Zimmermann and Church, 2001; Mueller et al., 2005; Gregoretti, 2008; Scheingross et al., 2013) and more recent theory (e.g., Lamb et al., 2008; Recking, 2009) suggest the opposite, however: sediment transport is less efficient in steep channels, as compared to lowland rivers, possibly due to bedforms such as step pools (Fig. DR1 in the GSA Data Repository[1]), changes in the hydrodynamics of shallow, rough flows, or incomplete submergence of grains during transport.

Field observations indicate that debris flows can dominate bedrock incision in very steep channels ($\theta > 6°$) and control the supply of sediment to channels downstream (e.g., Benda et al., 2005) (Fig. DR1). For example, topographic analyses indicate that the power-law scaling between channel slope and drainage area expected for river incision does not exist at very steep slopes, with the transition occurring at $\theta \approx 6°–35°$ in different landscapes (e.g., DiBiase et al., 2012), potentially signifying the onset of debris-flow transport (Stock and Dietrich, 2003). Debris flows can be triggered on hillslopes from shallow landslides (e.g., Iverson et al., 1997) or within channels due to bulking and failure of the channel bed (e.g., Takahashi, 1978; Gregoretti, 2000; Tognacca et al., 2000; Coe et al., 2008). The latter mechanism must control initial sediment motion in channels steeper than a critical slope, but this slope has yet to be identified. Herein we show results from exploratory experiments designed to identify the onset of sediment motion for a range of steep channel slopes that cross the river to debris-flow transition.

METHODS

We conducted 44 experiments in a 5-m-long tilting flume (Fig. DR2) with variable channel widths (35 and 13 cm) and 19 bed slopes ranging from $\theta = 1.8°$ to $33°$ (Table DR1 in the Data Repository). In natural channels, mixed sediment sizes and bedforms can influence initial sediment motion, and both these effects have been explored previously (e.g., Parker et al., 1982; Zimmermann et al., 2010). Here we focus on isolating the effect of channel-bed slope on initial sediment motion by using a planar bed of natural, well-sorted, semi-angular river gravel with a median intermediate grain diameter (D) of 1.5 cm. The grain size was chosen to achieve a sufficiently high particle Reynolds number,

$$Re_p \equiv \left(\frac{\tau}{\rho}\right)^{1/2} \frac{D}{\nu}, \tag{1}$$

where $\tau = \rho g H \sin\theta$ is the spatially and temporally averaged basal shear stress from surface flow, H is flow depth, g is gravitational acceleration, $\rho = 1000$ kg/m^3 is water density, and ν is the kinematic viscosity of water, such that viscosity and particle size do not affect initial motion (Shields, 1936). All experiments were repeated 2–4 times to assess error and natural variability.

We incrementally increased the water discharge (by 5%–15%; measured using a flow meter), pausing for 3–5 min at each discharge for measurements of sediment transport. In experiments with fluvial transport, we measured the volumetric sediment flux per unit width, q_b, for 1–3 min using a trap (Fig. DR2). As in previous work (e.g., Parker et al., 1982), the dimensionless bedload flux,

$$q_b^* = \frac{q_b}{\sqrt{\frac{\rho_s - \rho}{\rho} g D^3}}, \tag{2}$$

was a nonlinear function of the dimensionless bed stress, or Shields stress,

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

where $\rho_s = 2650$ kg/m^3 is sediment density. The critical Shields stress at initial motion, τ_c^*, was calculated by interpolating a power-law fit between τ^* and q_b^* to a standard reference transport rate of $q_b^* = 6.3 \times 10^{-5}$ corresponding to near initial-motion conditions (Parker et al., 1982) (Fig. 1). Average flow velocity was calculated by tracking pulses of dye. Non-Darcian

[1]GSA Data Repository item 2014067, Table DR1 (experimental findings), Note DR1 (note on the friction angles), Note DR2 (note on grain velocity measurements), Movies DR1–DR4, and Figures DR1–DR4, is available online at www.geosociety.org/pubs/ft2014.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

GEOLOGY Data Repository item 2014067 | doi:10.1130/G34927.1

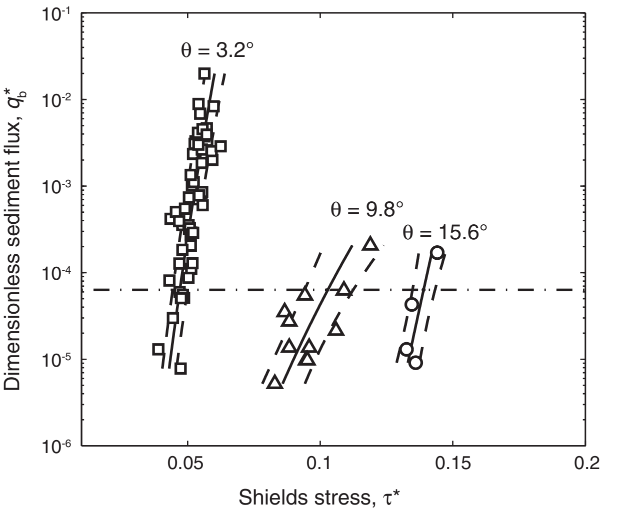

Figure 1. Dimensionless sediment flux (q_b^*) for experiments with fluvial transport at three different slopes (θ) as function of dimensionless shear stress (Shields stress, τ*). Solid lines are power-law fits and sub-parallel dashed lines represent 50% confidence limits. Critical Shields stress for each slope is defined at a reference dimensionless sediment flux for near-initial motion conditions (Parker et al., 1982) (horizontal dashed line).

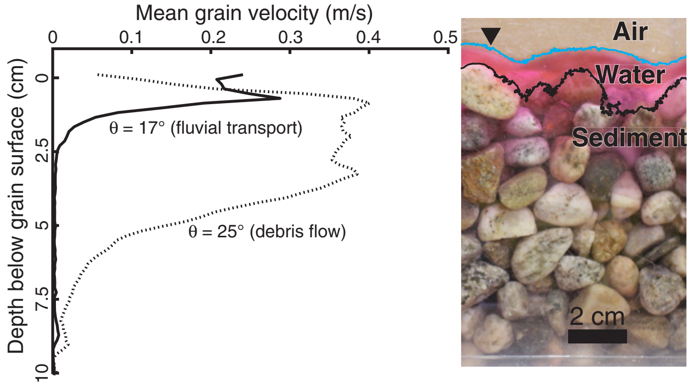

Figure 2. Grain velocity profiles in experiment with fluvial sediment transport (channel-bed angle θ = 17°) and mass failure of bed (θ = 25°), and side-view photograph of experimental bed with same vertical scale. Grain velocities were calculated using an autocorrelation routine applied to side-view videos (Leprince et al., 2007) (Note DR2 and Movies DR1 and DR2 [see footnote 1]). Grain velocities for fluvial case and at initiation of bed failure exist only within first grain layer (~1.5 cm). Bed failures developed into debris flows and entrained underlying grains, to depth of about five grains (~7.5 cm) for example shown. Photograph shows air-water (blue) and water-gravel (black) interfaces, which we identified semi-automatically with user-defined threshold value for each image, that were used to calculate flow depths in some experiments (Table DR1 [see footnote 1]).

subsurface discharge was calculated using a calibrated Forchheimer relation (Forchheimer, 1901) (Fig. DR3). Flow depth (*H*) was calculated by one or both of the following methods (see Table DR1): mapping and differencing sediment-water and water-air interfaces in side-view photographs at two locations (Fig. 2); or using continuity, flow velocity, and surface-flow discharge (i.e., differencing fully saturated subsurface discharge from total water discharge). For cases in which both methods were used, the difference between the two flow-depth methods was less than 30%, error that is smaller than variance between repeat experiments. There was no sediment feed, and experiments were ceased when sediment transport significantly altered the bed surface from its initial planar configuration to avoid the influence of bedforms.

To compare results to theory for fluvial transport (e.g., Wiberg and Smith, 1987; Lamb et al., 2008), we made 296 measurements of the inclination angle at which a single dry grain rolls from a fixed bed of similar grains (i.e., a grain-pocket friction angle of $\phi_g = 58.8° \pm 13.7°$, s.d., e.g., Miller and Byrne, 1966; see Note DR1 in the Data Repository). To apply the Takahashi (1978) bed-failure model, we performed 20 additional tilt-table measurements of the inclination angle (or failure-plane friction angle, ϕ_f) required to destabilize a collection of loose, dry grains. The bulk angle of repose for many grains using a number of different methods was measured as 45.6° ± 1.6° (s.d.; Note DR1). In our experiments, however, we observed that bed failures often initiated as a patch of loose grains ~7 grains long and ~1 grain deep, consistent with the Takahashi (1978) model for shallow bed failure in the presence of surface flow. Using the tilt table, we found that the failure-plane friction angle increased with decreasing number of loose grains, presumably due to stability derived from particle force chains (Cates et al., 1998), and that a failure-plane friction angle of $\phi_f = 55°$ is appropriate for the size of bed failures we observed (Note DR1 and Fig. DR4).

RESULTS

Initial sediment motion for experiments with 1.8° < θ < 19.6° occurred by river processes in which individual particles rolled and bounced along the bed driven by dilute surface flow (Fig. 2; Movie DR1 in the Data Repository). Although sediment transport is typically assumed to be a function of the Shields stress (τ*) only (e.g., Parker et al., 1982), our data show a clear shift to larger Shields stresses with steeper channel slopes for the same sediment flux (Fig. 1). Results show a marked increase in the critical Shields stress with channel slope from values typical of lowland rivers (e.g., $\tau_c^* = 0.035$ at θ = 1.8°; Buffington and Montgomery, 1997) to a value nearly 5 times as large at steep slopes (e.g., $\tau_c^* = 0.16$ at θ = 19.6°; Fig. 3). Thus, counter-intuitively, sediment transport by river processes requires larger bed stresses on steeper slopes despite the increased component of gravity acting on the grains in the downslope direction.

In experiments with θ > 25.2°, initial sediment motion occurred by mass failure of the bed (Fig. 2; Movies DR2–DR4) prior to any fluvial sediment transport. For these cases, collections of many grains mobilized together with the initial failure plane occurring within one or two grain diameters of the bed surface, and extending ~10 cm in the streamwise direction (Movie DR4). Bed failures mobilized into debris flows in which sediment and water were well mixed, and the dilute surface flow present at initial bed failure was mixed into the slurry (Movies DR2 and DR3). The debris flows often developed well-defined frontal snouts, and at very steep slopes would entrain sediment and run out through the end of the flume, and at lesser slopes would thicken, dewater, and stabilize. Unlike the fluvial transport regime, the Shields stress at initial sediment motion by bed failure decreased with increasing channel-bed slope from its maximum value of $\tau_c^* = 0.24$ at θ = 25.2° to $\tau_c^* = 0.053$ at θ = 33.0°, indicating a rapid decline in bed stability at steeper slopes (Fig. 3).

Experiments at θ = 22.3° exhibited both transport behaviors, with fluvial sediment transport occurring in some parts of the flume and small failures occurring elsewhere, representing a mixed transport regime. Therefore, θ ≈ 22° is the threshold slope in our experiments for the transition from initial motion by river to debris-flow transport.

Flume data for θ < 19.6° are consistent with the models of Lamb et al. (2008) and Recking (2009) that predict increased τ_c^* with increasing bed slope due to changes in surface-flow hydrodynamics and partially submerged grains in shallow, rough flows, and are inconsistent with models that do not consider these effects (Wiberg and Smith, 1987) (Fig. 3). Rather than changes in hydrodynamics, other workers have attributed enhanced sediment stability in steep channels to interlocking across the

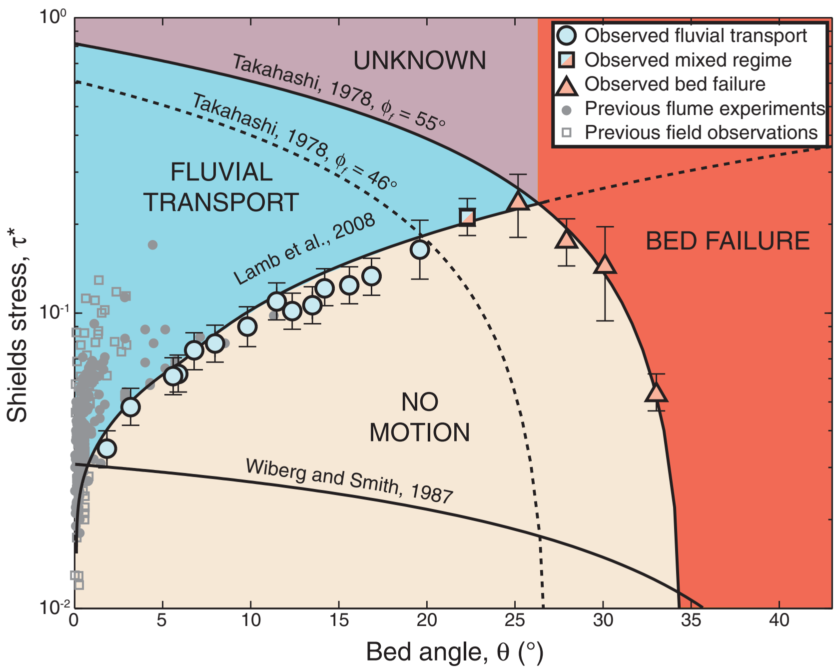

Figure 3. Experimental data of Shields stress at initial sediment motion as function of bed angle for both fluvial transport and bed failure. Error bars are 50% confidence limits for fluvial transport, and range of observed values for repeat bed failure experiments. Also shown are data from previous compilations (Lamb et al., 2008; Buffington and Montgomery, 1997) of initial sediment motion by fluvial transport including both field and laboratory observations. Curves are model predictions for incipient fluvial transport by Lamb et al. (2008) (dotted line shows model prediction within observed bed failure regime) and Wiberg and Smith (1987), and bed-failure model of Takahashi (1978) using failure-plane friction angle (ϕ_f) equal to bulk angle of repose (dashed line) and equal to what we measured for a loose patch of grains, ϕ_f = 55° (Note DR1 [see footnote 1]; solid line). Data support division of parameter space into conditions that produce no sediment motion, fluvial sediment transport, and bed failure. Data are not available to test mode of transport at shallow slopes and very large Shields stresses within regime marked "unknown."

channel width or changes in bed morphology (e.g., step-pool bedforms) (e.g., Zimmermann et al., 2010), or the presence of large immobile grains that increase flow resistance (Yager et al., 2007). These mechanisms cannot explain our results because our experiments had loose, planar beds of uniform grain sizes, and we observed no influence on incipient motion with changing channel width (Table DR1).

Despite good agreement with the fluvial-sediment transport model for θ < 19.6, Lamb et al. (2008) predict heightened τ_c^* values up to 0.4 at slopes that approach the grain-pocket friction angle of ϕ_g = 58.8°, which was not the case in our experiments due to the transition to bed failure at θ ≈ 22°. Instead, we compare the bed failure data to an infinite-slope, Mohr-Coulomb stability model in which the gravitational body force acting downslope on the groundwater and surface flow are balanced by frictional stress borne by the sediment grains (Takahashi, 1978) (Fig. 3). The model of Takahashi (1978) can be recast in terms of a critical Shields stress,

$$\tau_c^* = (1-\eta)(\tan\phi_f - \tan\theta) - \frac{\rho}{\rho_s - \rho}\tan\theta, \qquad (4)$$

where η is porosity, and we find that the model under-predicts our data when the failure-plane friction angle (ϕ_f) is set to the angle of repose (ϕ_f = 46°). Alternatively, the model matches the data using a larger failure-plane friction angle (ϕ_f = 55°), which is consistent with our measurements for a collection of loose grains of similar number to the observed failures (Note DR1; Fig. 3). It is also possible that the offset between data and model is due to turbulent subsurface flow, although the momentum balance of Takahashi (1978) does not require an explicit assumption of Darcian flow.

DISCUSSION AND CONCLUSIONS

Our results suggest that the threshold slope for initial sediment motion by debris-flow transport can be calculated by setting equal the transport models of Lamb et al. (2008) and Takahashi (1978) for specific failure-plane and grain-pocket friction angles. Given natural variability in these two friction angles [e.g., ϕ_f ~ 40° to 60° (Selby, 1993, p. 354); ϕ_g ~ 50° to 70° (Miller and Byrne, 1966)], we expect the transitional slope in natural channels to range from θ ≈ 15° to 30° (Fig. 4A), which is consistent with the observed change in slope–drainage area relationships at these slopes (e.g., Stock and Dietrich, 2003; DiBiase et al., 2012). As in our experiments, failure plane friction angles may be a function of failure size for coarse-grained natural channels due to the effects of particle force chains (e.g., Cates et al., 1998). The transition from fluvial to debris-flow transport can also be affected by drag due to flow separation behind bedforms or immobile boulders (i.e., form drag) in natural channels (Fig. DR1). For example, an increase in form drag from 40% to 90% of the total bed stress corresponds to a decrease in the calculated transitional slope from θ = 22° to 13° (Fig. 4B). Thus, the propensity for in-channel debris-flow initiation on steep slopes results from both the reduced effectiveness of fluvial transport and the increased effectiveness of subsurface and surface flow in destabilizing the sediment bed en masse.

The threshold slope for the transition to bed failure defines the steepest-sloping channels in which river processes play a role in sediment transport and channel incision. Channels with θ > 25° are typically devoid of

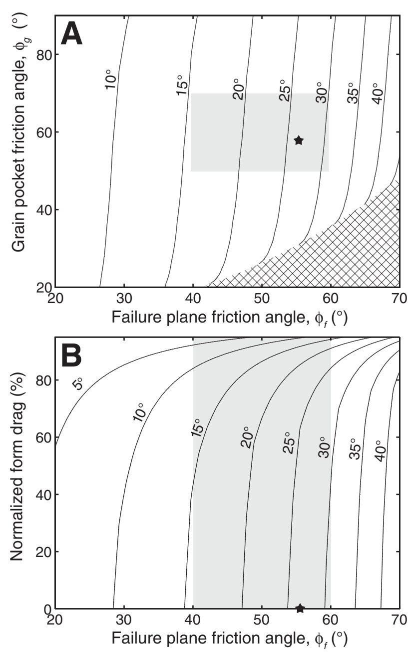

Figure 4. Contours of transitional slope between initial motion by fluvial and bed failure as calculated by matching models of Lamb et al. (2008) and Takahashi (1978) (e.g., Fig. 3) as function of grain-pocket friction angle (ϕ_g) and failure-plane friction angle (ϕ_f) with no drag due to bedforms (A), and as function of percentage form drag due to bedforms relative to total bed stress and failure-plane friction angle with constant 58.8° grain-pocket friction angle (B). Shaded regions denote range of parameter space likely in natural channels and black stars denote our experiments. Crosshatched region of A is model parameter space for dry bed failure.

river-sorted sediment, and instead contain poorly sorted colluvium, rockfall, or debris-flow deposits (e.g., Fig. DR1). This notwithstanding, debris flows influence channel processes at lesser slopes ($\theta < 25°$) because, even if initiated within steep channel reaches, they can run out long distances and across channels with slopes as little as $\theta = 1°$ (e.g., Iverson, 1997). In addition, Figure 3 suggests that mass failure of channel beds can occur at slopes lower than the transitional slope if Shields stresses during floods surpass substantially the critical value for fluvial transport (i.e., $\tau^* \rightarrow 1$). Shields stresses that substantially exceed the threshold for motion are rare in coarse-bedded rivers (e.g., Parker et al., 2007), but may occur due to infrequent, large-magnitude floods (which increase τ^* by increasing τ), or due to abrupt pulses of fine sediment (which increase τ^* by decreasing D). The latter mechanism may partly explain the propensity for debris flows (or debris floods) following landslides or wildfire (e.g., Coe et al., 2008; Lamb et al., 2011), for example, where hillslope-derived pulses of sediment can effectively resurface channels with finer sediment.

A quantitative framework for initial sediment motion across the river–debris flow transition allows for more realistic treatment of erosion and sediment transport in landscape-evolution models and hazard-mitigation efforts by providing process partitions in the landscape. Given that channels typically adjust their geometries to pass sediment at bed stresses that just exceed that required for initial sediment motion (e.g., Parker et al., 2007), the fivefold increase in the critical Shields stress near the transitional slope, as compared to the value typically assumed based on lowland river studies, provides new expectations of river form and may explain why sediment-transport models drastically over-predict sediment flux in steep channels (e.g., Rickenmann, 1997). Finally, our study adds to increased recognition of the role, in landscape evolution and mountain hazards, of debris flows initiated in-channel rather than on hillslopes (e.g., Takahashi, 1978; Gregoretti, 2000; Tognacca et al., 2000; Coe et al., 2008).

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