APPENDIX G

Evaluation of Sediment Transport Conditions

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Shear Stress Calculation

Calculation of the stream's bed shear stress (τ_0), or tangential force per unit bed area, is necessary to understand flow intensity and its ability to mobilize and transport sediment particles resting on the bed. Bedload transport rates (e.g., gravel transport) are steep and non-linear, which means relatively small changes in shear stress can create large changes in sediment transport. Therefore, obtaining accurate shear stress estimates is critical in calculating sediment transport.

For steady, uniform flow the momentum equation states a balance must exist between shear forces (resisting forces) and gravity component (driving forces)

$$\tau_0 P_w \Delta s = \rho g A \Delta s S$$

or

$$\tau_0 = \rho g R S$$

where τ_0 is bottom shear stress, P_w is wetted perimeter, Δs is length of control volume, ρ is fluid density, g is gravity acceleration, A is cross-section area, S is the bed slope, and R is the hydraulic radius.

To calculate bed shear stress for steady, gradually varied flow conditions common to most streams, the friction slope S_f is often substituted for the bed slope S. And for relatively wide channels where the hydraulic radius and mean flow depth are approximately similar, the "depth*slope" product is used to calculate the mean cross-sectionally averaged boundary shear stress

$$\tau_0 = \rho g H S_f$$

where H is mean flow depth.

The mean boundary shear stress, however, includes forces acting on debris jams, vegetation, channel banks, bar forms, and other features that add resistance and increase flow depth. Research has shown that the actual bed shear stress available for sediment transport (effective shear stress) is often a third to a half the mean boundary shear stress (Dietrich 1987). To gain a better estimate of only the portion of the shear stress that is acting on the sediment grains and available to transport sediment, a local estimate of shear stress directly above the area of the bed of interest is required. This local estimate is often referred to as a grain stress.

The following section describes the method used in this study to calculate local bed shear stress.

Time averaged fluid shear stress in a streamflow is defined as the rate of change of downstream momentum per unit cross-sectional area

$$\tau = -\rho \overline{u'v'}$$

where τ is turbulent shear stress, ρ is fluid density, u' is downstream velocity, and v' is vertical velocity.

Determining the vertical variation in flow velocity in turbulent flow requires knowledge of the mixing length l, or the vertical distance over which a fluid parcel's momentum changes. By equating the mixing length to

$$u' = -l\left(\frac{d\overline{u}}{dy}\right) \quad \text{and} \quad v' = l\left(\frac{d\overline{u}}{dy}\right)$$

then turbulent shear stress is

$$\tau = -\rho \overline{u'v'} = \rho l^2 \left(\frac{d\overline{u}}{dy}\right)^2$$

By assuming that: (1) the fluid shear is approximately equal to the bed shear near the streambed; and (2) mixing length increases linearly with distance from the bed, the law of the wall equation for determining the velocity gradient near the streambed (i.e., "wall") is calculated from

$$\frac{\overline{u}}{u*} = \frac{1}{\kappa}\ln\left(\frac{y}{y_0}\right)$$

where κ is Von Karman's constant (commonly set at 0.41), $\overline{u}$ is time averaged velocity at flow depth y above the bed, and y_0 is the flow depth where flow velocity equals zero. The shear velocity, $u*$, is a measure of the velocity gradient near the bed, from which local bed shear stress can be calculated

$$\tau_0 = u*^2 \rho$$

In reality, flow velocity is only zero where $y = 0$. Therefore, in order to solve the equation for hydraulically rough flows, y_0 is related to the equivalent roughness height, k_s, by

$$y_0 = \frac{k_s}{30}$$

And k_s is based on the dominant coarse bed substrate, such as the D_{84} (the particle size in which 84% of the bed surface is finer).

Integration of the law of the wall equation above over the entire flow depth (h) shows that the mean flow velocity occurs at a distance of $0.368h$ from the bed. By inserting the $0.368h$ and k_s values into the law of the wall equation above, the local shear velocity, and thus local shear stress related to grain-induced resistance can be determined from mean channel velocity (U) using Keulegan's (1938) resistance law for rough flow:

$$\frac{U}{u*} = \frac{1}{\kappa}\ln\left(\frac{h}{k_s}\right) + 6$$

This equation, or variations of it, is commonly used to calculate local shear stress values for use in incipient motion and sediment transport analysis.

The following equation (from Wilcock 1996) was used to calculate local grain stress in this study:

$$\frac{U}{u*} = \frac{1}{\kappa}\ln\left(\frac{h}{ez_0}\right)$$

where z_0 (the bed roughness length where flow velocity (u) is 0) is calculated from

$$z_o = \frac{3D_{84}}{30}$$

Initiation of Motion Calculation

Whether or not a particle on the stream bed will be entrained by the flow or remain in place depends on: (1) randomness (grain placement and turbulence); and (2) balance of driving fluid drag (F_D) and resisting gravity forces (F_G)

$$F_D \propto \tau_0 D^2, \text{ and } F_G \propto (\rho_s - \rho)gD^3$$

and

$$\frac{F_D}{F_G} \propto \frac{\tau_0}{(\rho_s - \rho)gD} = \Theta = \tau^*$$

where D is grain diameter and ρ_s is sediment density. The dimensionless bed shear stress (Θ, commonly called the Shields number, or τ^*) is a measure of sediment mobility. If τ^* is greater than the threshold required for sediment motion (τ^*_c, critical dimensionless bed shear stress), then sediment motion is predicted to occur.
Much research continues in the field of sediment movement initiation and the selection of appropriate τ^*_c. Figure G-1 shows initiation of motion curves from which τ^*_c is

determined from the particle Reynolds number (R_{ep}). If the τ^*_c value plots above the curve, then sediment motion is predicted to occur; whereas if the value is under the curve, then no motion is predicted to occur. Both curves show that as particle size increases from coarse sand to gravel, the increased resistance to movement from the weight of the particle exceeds the additional drag exerted on the particle, and thus the critical dimensionless shear stress required for movement increases. The curves flatten out at as particle size approaches coarse gravel (32-64 mm) and coarser particles. Several researchers have shown the original Shields curve (in blue) values for initiation of motion are too high, and thus predict too much shear stress is required for sediment movement. Therefore, Figure G-1 shows a modified curve (in red) in which the initiation of motion curve flattens out around 0.045 instead of 0.06.

The same two original and modified Shields curves are plotted in dimensional units in Figure G-2. From this plot, the amount of shear stress (Pascal units) needed to initiate motion of a given particle size (mm units) can be determined.

Initiation of Motion for Sediment Mixtures

The initiation of motion curves in Figures G-1 and G-2 represent critical shear stress values needed to mobilize sediment of a uniform size resting on a nearly flat channel bed. The curves do not consider how the relative variability of grain sizes in a sediment mixture influence initiation of motion values for individual particle sizes (D_i) within the mixture. For sediment mixtures of coarse and fine particles, the coarser particles (e.g., gravel) in the mixture can be relatively easier to mobilize than if all the sediment was the same size because the coarser grains protrude higher into the flow where flow velocities are greater, and they have relatively lower pivoting angles. By contrast, the smaller particles in the sediment mixture have higher pivot angles, and are shielded from the higher flow velocities by the larger particles. Therefore, the finer (e.g., sand) particles in a mixture can be relatively harder to mobilize than if all the sediment was the same size.

Additionally, research has shown the importance of the percentage of sand in a sediment mixture on the critical shear stress needed to mobilize both sand and gravel particles (Wilcock 1998; Wilcock and Crowe 2003). As the sand content increases on the bed to larger percentages, the gravel particles become less constrained by other gravel particles, and thus more of the particle is exposed to fluid drag since it is becoming larger than its surroundings. Once the gravel particle is entrained, it moves faster over the relatively smooth, mobile bed created by the sand than it would over other gravel particles, and it may move a greater distance because potential resting areas are filled with sand. At even higher percentages of sand, gravel particles can be mobilized through undercutting of the underlying sand, and once mobilized the gravel keeps going over the relatively smooth sand bed. Figure G-3[1] shows how variations in bed surface sand content influence the critical dimensionless shear stress needed to

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[1] The reference shear stress values presented in Wilcock and Crowe (2003) were converted to critical shear stress values by reducing the reference shear stress by 10%, per Wilcock 1998.

initiate motion of a sediment mixtures mean particle size (D_m) (Wilcock and Crowe 2003). Figure G-4 is the same plot but with dimensional critical shear stress values for different D_m values. The plots show that as surface sand content increases from 0 to 20%, the shear stress needed to mobilize the D_m decreases. Sand content increases greater than 20% have little influence on the critical shear stress needed for sediment initiation.

The Wilcock and Crowe (2003) method for calculating the critical shear stress needed to initiate sediment movement for mixed-size sediment was used for this study. This method was chosen since it considers how relative particle size variation within the sediment mixture and sand content influence sediment mobility.

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TABLES

FIGURES