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Manual for Computing Bed Load Transport Using BAGS (Bedload Assessment for Gravel-bed Streams) Software

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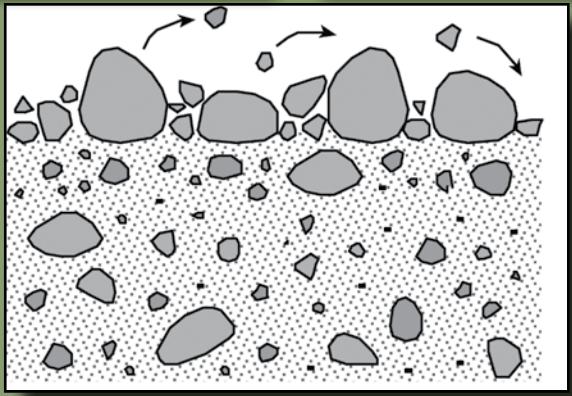

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Abstract

This manual provides background information and instructions on the use of a spreadsheet-based program for Bedload Assessment in Gravel-bed Streams (BAGS). The program implements six bed load transport equations developed specifically for gravel-bed rivers. Transport capacities are calculated on the basis of field measurements of channel geometry, reach-average slope, and bed material grain size. Calculations are carried out using Visual Basic for Applications (VBA), and the output is stored on individual spreadsheets. In addition to step-by-step instructions in software operation, the manual provides guidance in the interpretation of results.

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Download Information

The BAGS program, this manual, and a sediment transport primer (Wilcock and others 2009) can be downloaded from: http://www.stream.fs.fed.us/publications/software.html.

This publication may be updated as features and modeling capabilities are added to the program. Users may wish to periodically check the download site for the latest updates.

BAGS is supported by, and limited technical support is available from, the U.S. Forest Service, Watershed, Fish, Wildlife, Air, & Rare Plants Staff, Streams Systems Technology Center, Fort Collins, CO. The preferred method of contact for obtaining support is to send an e-mail to rmrs_stream@fs.fed.us requesting “BAGS Support” in the subject line.

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Contents

Background on Data Input and Software Operation

This manual for computing bed load transport provides specific instructions on the use of BAGS (Bedload Assessment of Gravel-bed Streams) software with worked examples illustrating typical results and comments on possible interpretations of the output. The BAGS software calculates bed load transport capacities on the basis of commonly available field data (surveyed cross sections and measured grain size distributions) and stores the output as tabulated values on individual spreadsheets. These values can be retained for further analyses or exported to other programs for plotting and visualization. Options within the program allow the user to select from six transport relations developed specifically for gravel-bed streams and rivers. Transport rates can be calculated for a single flow or a series of flows, depending on the application and data availability. The first section of this manual gives an overview of the basic data requirements and introduces some important concepts shared by each of the bed load transport relations—<u>all users should read this section</u>! Details of the transport relations and their various components are presented in the second section. The third and fourth sections describe the steps involved in running the software. The final section provides several examples illustrating results from different model runs and discusses problems that might arise in typical applications. This manual is a companion to "Sediment Transport Primer: Estimating Bed-Material Transport in Gravel-bed Rivers" (Wilcock and others 2009).

The Sediment System

Each of the models used in calculating bed load transport rates has some common data input requirements, these include:

- a measured channel cross section (or, at a minimum, an estimate of the bankfull width);
- an estimate of the reach-average slope, obtained from measurements of water-surface elevations or bed elevations;
- discharge measurements (and a flow duration curve if one of the goals is to estimate the long-term annual bed load sediment yield); and
- grain size parameters estimated from samples of the bed sediment.

The primary differences in software operation and data inputs center around sediment properties and various measures of these properties. In gravel-bed streams, we can separate the "sediment system" into three distinct components, as illustrated in figure 1. The sediment we typically see on the **bed surface** consists of relatively coarse clasts representing the largest grain sizes carried by the

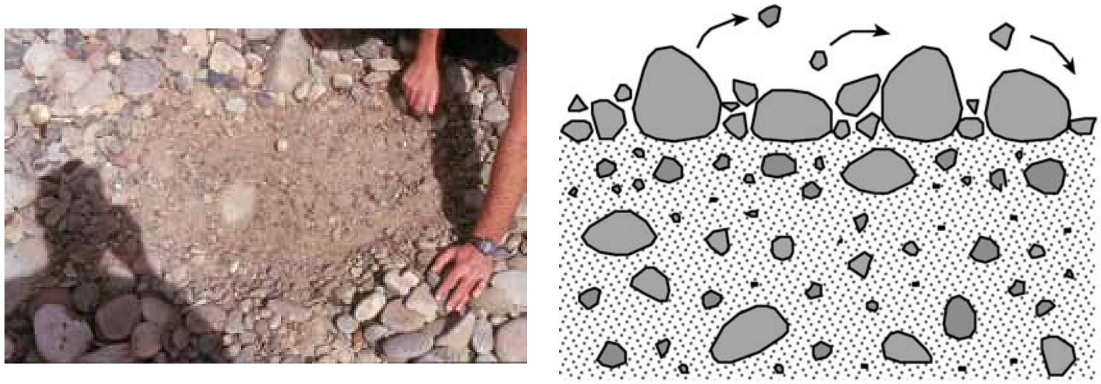


Figure 1. Distinction between the bed surface layer (armor) and the substrate, as commonly found in gravel-bed streams and rivers.

stream or river. Winnowing of finer particles from the bed surface produces a distinct layer with a thickness approximately equal to the diameter of the coarsest grains. This sediment is referred to as the **surface layer** or armor. The sediment immediately beneath the surface layer consists of a more homogeneous mixture of fine and coarse particles and is referred to as the **substrate** or bulk bed material. The substrate is generally much finer than the surface layer, and typically 10 to 20 percent of this sediment is finer than gravel (<2 mm). The third component of the sediment system is the **bed load** itself. This sediment is, by definition, the material that moves in contact with the bed.

We can illustrate the differences in sediment characteristics graphically by plotting the distribution of particle sizes as cumulative frequency curves. Figure 2 presents two examples of grain size distributions based on sediment samples taken in two gravel-bed rivers. Both data sets include measurements of the bed load, bed surface layer, and substrate. The panel on the left shows that the bed load carried by the Salmon River is much finer than the substrate—the

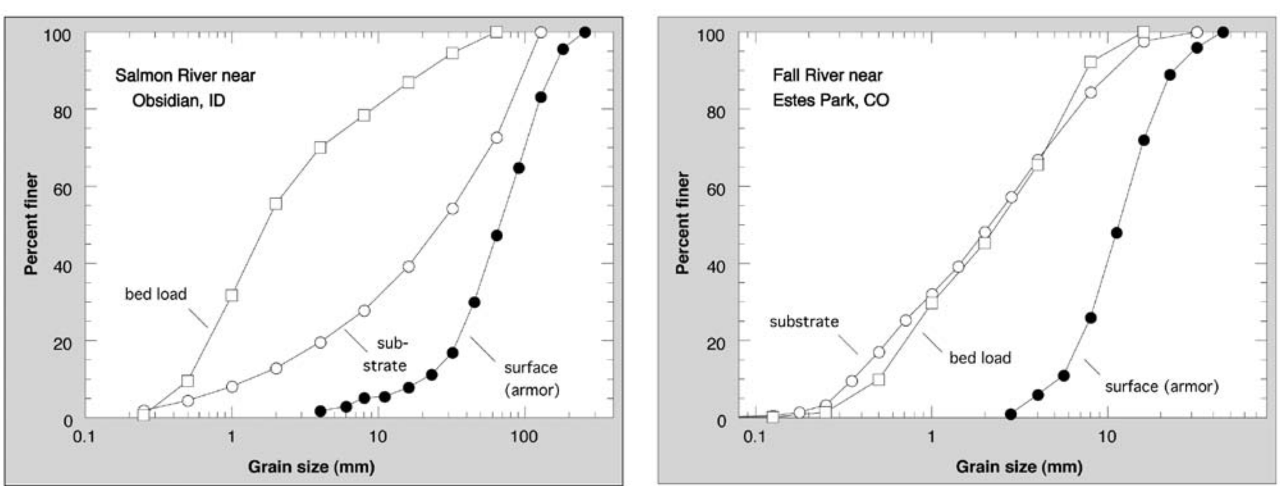


Figure 2. Grain size distributions of the bed load, surface layer, and substrate in two gravel bed rivers. Salmon River data obtained through the Boise Adjudication Team web site (http://www.fs.fed.us/rm/boise/research/watershed/BAT). Fall River data from Pitlick (1993).

substrate, in turn, is much finer than the bed surface. In this case, the three components of the sediment system are quite distinct from each other and it does not appear that much of the load is derived from the bed. In contrast, the panel on the right shows that the bed load carried by Fall River has nearly the same size distribution as the substrate. The similarity in bed load and substrate grain size distributions suggests a clear link between the load and the source: in this case, it appears that the bed load and substrate grain sizes exchange with each other almost on a one-for-one basis.

The examples shown in figure 2 are more or less representative of the range of conditions one is likely to encounter in measuring or modeling transport rates in gravel-bed streams. However, it would be difficult to know in advance whether a given stream in a specific geographic location should act more like the Salmon River or Fall River, and it would be impossible to determine this without extensive sampling. Results from field studies and flume experiments suggest that the separation between bed load and substrate grain-size distributions is related to variations in hydrology, boundary shear stress, and/or sediment supply, although these interactions are not very well understood (Barry and others 2004; Buffington and Montgomery 1999; Dietrich and others 1989; Hassan and Church 2001; Hassan and others 2006; Lisle 1995; Mueller and others 2005; Parker 1990a; Pitlick and others 2008; Powell and others 2001; Wathen and others 1995). Ultimately, all of the sediment carried by a river is supplied by the watershed; however, over the short time scales of individual flow events, some of the bed load may be derived from the bed itself and some from sources outside the channel (for example, stream banks or hillslopes). Success in predicting bed load transport rates hinges to a large extent on the availability of mobile sediment sizes within the channel boundary, but it is not always obvious that conditions at a particular location do or do not satisfy this requirement.

Overview of BAGS Software Operation

The BAGS software described in this manual implements six well-known bed load transport equations developed specifically for gravel-bed rivers:

- the surface-based equation of Parker (1990);
- the substrate-based equation of Parker-Klingeman-McLean (Parker and others 1982);
- the substrate-based equation of Parker and Klingeman (1982);
- the surface-based two-fraction equation of Wilcock (2001);
- the surface-based equation of Wilcock and Crowe (2003); and

- the procedure of Bakke and others (1999), which calibrates two coefficients in the substrate-based equation of Parker and Klingeman (1982).

All of the equations and procedures recognize the role of the armor layer in regulating bed load transport rates, thus the dynamics of transport are represented by the three components discussed above: the surface layer, substrate, and bed load. The equations of Parker (1990a), Wilcock (2001), and Wilcock and Crowe (2003) apply surface grain size characteristics as inputs, while the Parker-Klingeman-McLean equation (Parker and others 1982) uses substrate grain sizes. The method of Bakke and others (1999) applies to either the surface or the substrate, depending on circumstances. In addition, the methods of both Wilcock (2001) and Bakke and others (1982) use bed load measurements to calibrate certain coefficients in the transport equations and procedures.

The equations and procedures are implemented in an MS-Excel workbook with Visual Basic for Applications (VBA). Field data and relevant parameters are entered into the program sequentially with a series of user prompts. Results of the calculations are presented in MS-Excel workbooks. The software is designed to be used by professionals (hydraulic engineers, fluvial geomorphologists, and hydrologists) with some familiarity and training in processes of sediment transport. The correct interpretation of the results, however, typically requires a more in-depth knowledge of fluvial processes, especially the dynamics of transport in gravel-bed rivers.

Basic Data Input Requirements

Grain size

Representative samples of the bed surface layer or the substrate are required in order to develop a cumulative frequency distribution of available grain sizes. Sample values are entered into the program as a listing of the percentage of particles finer than a given size, D, where D is measured in millimeters. Different methods are used to sample the surface and the substrate. The surface layer is sampled by using some variation of the method introduced by Wolman (1954), or by taking a volumetric sample of the surface layer down to the depth of the largest clasts, as described by Milhous (1973). Substrate samples are taken by excavating a pre-determined mass of sediment from beneath the surface layer. There are many important issues to consider in sampling the surface layer and substrate, and we won't go into these details here. Procedures for sampling surface and substrate sediment and criteria for establishing sample sizes are described in detail in a number of papers and reports, including Kellerhals and Bray (1971), Church and others (1987), and Bunte and Abt (2001).

Figure 3 shows an example of the data listing for a point count of the surface layer and the resulting grain size distribution. The difference between the finest and coarsest grain sizes in a sediment sample is typically quite large, thus the grain size curve is plotted using a logarithmic scale for the x axis (fig. 3).

For convenience, a Ψ scale is also used to represent grain size, as shown on the secondary x axis. The Ψ scale varies with the base 2 logarithm of the grain size:

$$\Psi = \frac{\log D}{\log 2}\,,\ or\ D = 2^{\Psi} \qquad \text{(1a,b)}$$

where D is in mm. The Ψ scale used here is identical to the ϕ-scale used in sedimentology, except the sign is reversed, resulting in positive values of Ψ for D>1 mm. If the sample values are split at even increments of Ψ = 1, 2, 3… and so forth, then the grain size distribution is said to be sampled (or sieved) at 1-Ψ intervals. If higher resolution is required, sample values can be split at smaller intervals, for example, 1/2Ψ = 0.5, 1.0, 1.5, 2.0, … and so forth. The grain size distribution and data listed in figure 3 show measurements obtained at 1/2-Ψ intervals.

The software calculates transport rates and bed friction coefficients on the basis of discrete values of the grain size distribution, D_i, where the subscript i refers to an individual percentile of the grain size distribution. The midpoint of the distribution corresponding to the value for which half the sediment is finer is the median grain size, D_{50}. In the above example, D_{50} = 60 mm. Additional sizes referred to elsewhere in the manual include D_{65}, D_{84}, and D_{90}.

Other parameters are calculated from the full grain size distribution of the sample using individual values for each size class. Let D_1, D_2, …, D_{N+1} be the grain sizes associated with each of the N size classes, and let f_1, f_2, …, f_{N+1} be

Location: XS-10 **Date: 7/10/04**

D(mm)	Ψ	# passing	pct finer
512	9.0		100
362	8.5	1	100
256	8.0	5	99
181	7.5	10	94
128	7.0	8	84
90	6.5	23	76
64	6.0	21	53
45	5.5	14	32
32	5.0	12	18
22	4.5	5	6
16	4.0	1	1
11.2	3.5		
8.0	3.0		

total = 100

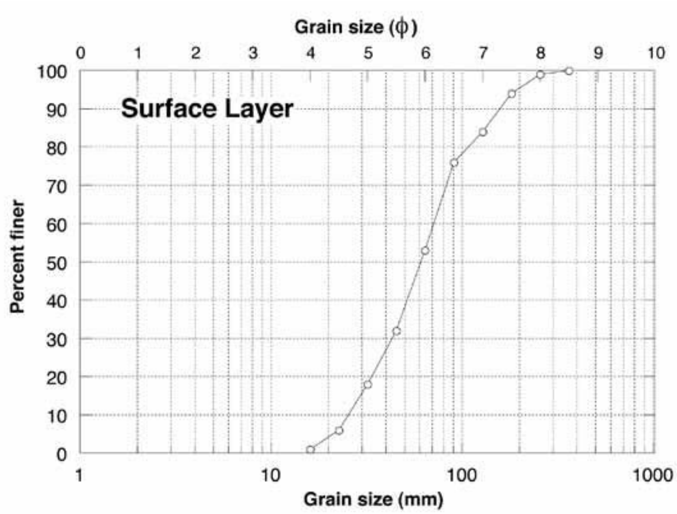


Figure 3. Listing of values obtained from point-count sample of the surface layer of a gravel-bed river, and the grain size distribution curve corresponding to this sample.

the fraction of the sampled mass (or the sampled number of grains) represented in each size class. The mean values of D_i, Ψ_i, and f_i for each class are calculated as follows:

$$\overline{D}_i = \sqrt{D_i D_{i+1}}, \quad \overline{\Psi}_i = \frac{\Psi_i + \Psi_{i+1}}{2}, \quad \overline{f}_i = f_{i+1} - f_i \qquad \text{(2a,b,c)}$$

where the subscripts i and $i + 1$ refer to adjacent size classes. The values obtained from (2) can be used to estimate additional parameters:

$$\overline{\Psi} = \sum_{i=1}^{N} \overline{\Psi}_i \, \overline{f}_i, \qquad D_g = 2^{\overline{\Psi}} \qquad \text{(3a,b)}$$

and

$$\sigma = \sqrt{\sum_{i-1}^{N} \left(\Psi_i - \overline{\Psi}^2 f_i \right)}, \qquad \sigma_g = 2^{\sigma} \qquad \text{(3c,d)}$$

where $\overline{\Psi}$ is the arithmetic mean (in Ψ units); D_g is the geometric mean grain size (in mm); σ is the arithmetic standard deviation (in $\overline{\Psi}$ units); and σ_g is the geometric standard deviation (in mm). In the preceding example, $\overline{\Psi} = 5.94$, $D_g = 60$ mm, $\sigma = 0.90$, and $\sigma_g = 1.9$ mm. The values of D_{50} and D_g are very nearly identical in this example; however, in general, D_g will be smaller than D_{50}.

Channel cross section

The software calculates relevant flow properties, such as the mean flow velocity and hydraulic radius, from a measured cross section. Depending on the application and site characteristics, you may want to include overbank areas (floodplains) in the calculations. The software will prompt you accordingly, and you will be asked to specify the left and right boundaries of the floodplain, as shown in figure 4. The hydrodynamic component of the model accounts for changes in the geometry and velocity of the flow as it spreads out across the

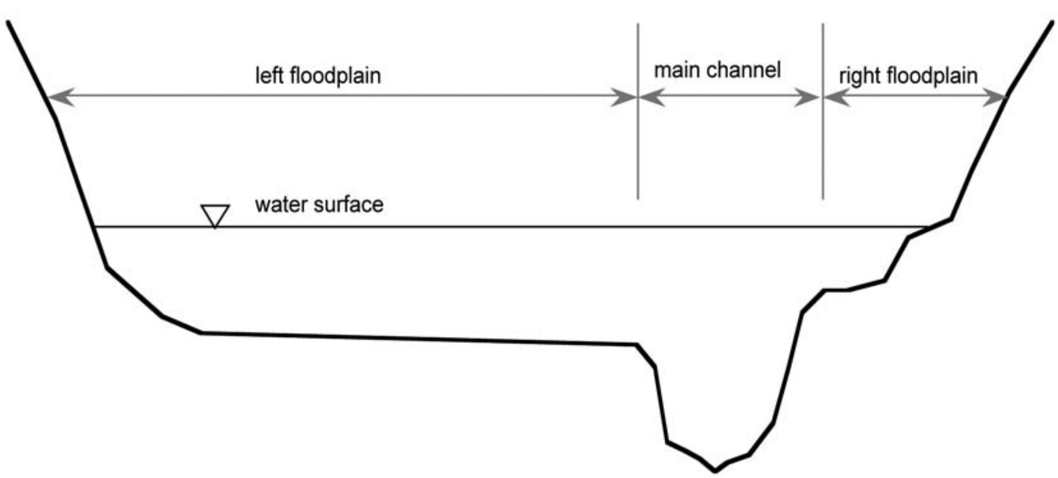


Figure 4. Sketch of a channel cross section. Water is assumed to flow through the main channel and the floodplains if inundated. Bed load is assumed to occur only in the main channel.

floodplain. However, the sediment transport component of the model restricts bed load calculations to the main channel under the assumption that the water-surface elevation is constant across the channel and the floodplain.

Flow resistance and shear stress

Flow resistance: The software estimates flow resistance in the channel separately from flow resistance over the floodplain. In the main channel, it is assumed that flow resistance is dominated by the stationary grains on the bed. Given the geometry of the channel and the grain size distribution of the bed surface, the software calculates flow properties using the Keulegan resistance relation:

$$\frac{U}{u_*} = 2.5 \ln\left(11 \frac{R_c}{k_s}\right) \tag{4}$$

where $u_* = (g\ R_c\ S_f)^{1/2}$ is the shear velocity; g is the gravitational acceleration; R_c is the hydraulic radius of the main channel, equal to the cross-sectional area, A_c, divided by the wetted perimeter, P_c; S_f is the friction slope; and k_s is the equivalent roughness. The convention followed here is to assume that k_s scales with a coarser-than-average grain size since larger grains contribute more to the total flow resistance than smaller grains. The criteria for choosing one value of k_s over another ($3D_{84}$, $2D_{90}$, and so forth) is based largely on empirical relations and individual preference. To be consistent with the original references, three slightly different values of k_s are used: Parker's surface-based equation assumes $k_s = 2D_{90}$; Parker's substrate-based equations assume $k_s = 10.7D_{50_{sub}}$, where the subscript *sub* refers to the substrate; and Wilcock's equations assume $k_s = 2D_{65}$.

Shear stress: The total shear stress acting on the bed and banks is calculated from the depth-slope product, $\tau_o = \rho g R S_f$, where ρ is the density of water (see Wilcock and others 2009, Chapter 2—Non-uniform and unsteady flow, for a more complete discussion of the components of shear stress). The proportion of the total shear stress available for transporting bed load—termed the "grain stress"—is estimated using a drag-partitioning algorithm that couples the Keulegan relation for velocity (4) with the Manning-Strickler relation for grain roughness (see Wilcock and others 2009, Chapter 2—The drag partition).

Overbank flows: Flow over the floodplain is characterized using a combined form of the continuity equation and the Manning equation:

$$Q_1 = \frac{1}{n_l} A_l R_l^{2/3} S_f^{1/2}\ , \qquad Q_r = \frac{1}{n_r} A_r R_r^{2/3} S_f^{1/2} \tag{5}$$

where Q is the discharge over the floodplain; n is a user-defined roughness coefficient; A is the cross-sectional area of the flow over the floodplain; R is the hydraulic radius; and the subscripts l and r refer to left and right portions of the floodplain, respectively. For overbank flows, the value of S_f is assumed to be the same across the channel and the floodplain. Ideally, estimates of S_f should be made with a one-dimensional hydraulic model, such as HEC-RAS or WS-PRO; however, if this information is not available, reach-average estimates of the water surface slope or the bed slope are used as approximations for S_f. Guidelines and techniques for estimating floodplain roughness are discussed in several publications, including Barnes (1967), Arcement and Schneider (1989), and Hicks and Mason (1998). Several web sites provide descriptions of techniques for estimating flow resistance in simple and compound channels and up-to-date references for published studies:

http://wwwrcamnl.wr.usgs.gov/sws/fieldmethods/Indirects/nvalues/

http://manningsn.sdsu.edu

http://www.rivers.gov.au/Tools_and_Techniques/Stream_Roughness_Coefficients_Tool/

Bed Load Transport Equations

This section provides a detailed discussion of the individual transport relations used in the BAGS model. We have tried to simplify the notation and terminology as much as possible while retaining key elements of each equation, as given in the original references.

The bed load transport relations used in the BAGS model are conceptually similar in that they all model transport rates as a function of the **transport stage**, which is simply a ratio of the available shear stress to the threshold shear stress:

$$\phi = \frac{\tau}{\tau_r} \tag{6}$$

where τ is defined as before and τ_r is the reference shear stress that produces a very small but measurable transport rate (Parker 1990; Parker and others 1982). This term is analogous to the critical shear stress, τ_c, which is used in a number of other bed load transport equations, including Meyer-Peter and Muller (1948) and Fernandez-Luque and van Beek (1976). The reference shear stress is used in place of the critical shear stress to avoid ambiguities in defining the transport threshold; bed load begins moving over a range of flows, and there is always some chance, even at low flows (ϕ<1.0), that a small number of grains might be moving (more on this later). It is also important to note that, as the ratio of τ to τ_r grows, there is a nonlinear response in transport, thus any errors associated with estimates of τ or τ_r are quickly amplified, leading to potentially large uncertainty in calculated transport rates.

Bed load transport rates are expressed in terms of a dimensionless parameter:

$$W^* = \frac{(s-1)gq_b}{\rho_s(\tau/\rho)^{1.5}} = \frac{(s-1)gq_b}{\rho_s u_*^3} \tag{7}$$

where s is the specific gravity of sediment; g is the gravitational acceleration; ρ_s is the density of sediment (2650 kg/m^3 or 2.65 g/cm^3 for quartz-density sediment); ρ is the density of water; u_* is the shear velocity; and q_b is the mass transport rate per unit width. The transport parameter W^* is likewise a ratio, in this case representing the power required to transport bed load scaled by the power available (Parker and others 1982). Values of W^* could thus be similar for a few large grains moving at high shear stress or many small grains moving at low shear stress. The primary reason for formulating the transport equations in terms of dimensionless parameters is to maintain generality so that the equations and results are transferable across a range of scales.

The other dimensionless parameter that appears frequently in the bed load transport equations is the **Shields stress**:

$$\tau^* = \frac{\tau}{(\rho_s - \rho)gD} = \frac{u_*^2}{(s-1)gD} = \frac{RS_f}{(s-1)D} \tag{8}$$

Eq. 8 is derived by balancing the drag force acting over the area of the grain against its weight. This equation, therefore, represents a ratio of the force available to move a given grain size versus the resistance provided by the weight and contact forces. Using this relation, Eq. 6 can be rewritten as:

$$\phi = \frac{\tau^*}{\tau_r^*} \tag{9}$$

where τ_r^* is termed the reference Shields stress.

For some applications, it may be necessary to compute transport rates for individual grain sizes, D_i, where the subscript i refers to the i-th size fraction of the grain size distribution. In this case, the parameters listed above are denoted ϕ_i , W_i^*, and $\tau_{r_i}^*$, and the equation will generally include additional terms representing the proportion of sediment in the i-th size fraction. The symbols, p_i , f_i , and F_i , are used to denote, respectively, the proportion of sizes in one of three potential populations of sediment: the bed load, substrate, and surface layer. Calculations of fractional transport rates also involve what is known as a "hiding function," a function that accounts for size-dependent differences in the mobility of small and large grains (Andrews 1994; Parker and others 1982; Wilcock and Crowe 2003). In gravel channels, small lightweight grains tend to get lodged in the interstices between large grains, hence they are hidden from the flow and less mobile than they might be otherwise. Large grains protrude into the flow and are exposed to higher velocities, hence they are more mobile than they might be otherwise. The offsetting effects of hiding and exposure are reflected in the hiding function by an inverse relation between the reference Shields stress for an individual grain size, $\tau^*_{r_i}$, and the ratio of the individual grain size to the median grain size, D_i/D_{50}.

Finally, several of transport relations used in the BAGS model consist of more than one function, with each function covering a different range in transport stage, ϕ (see the first set of equations below). There are two principal reasons for developing multi-part functions: (1) Bed load data often exhibit a slight but distinct curvature when transport rates are plotted against τ^* or ϕ on log-paper, thus a single power law equation may not fit the data particularly well across the entire range of values and (2) It has been shown that the transport rate, W^*, should approach a constant at large values of ϕ (Parker and Klingeman 1982; Yalin 1972), thus it is desirable to have one component of the transport function satisfying this condition.

Surface-Based Bed Load Equation of Parker (1990)

The surface-based bed load equation of Parker (1990a) consists of three matching functions representing different levels of transport intensity:

$$W_i^* = \begin{cases} 11.9\left(1 - \dfrac{0.853}{\phi}\right)^{4.5} & \phi_{50} > 1.59 \\ 0.00218\exp\left[14.2\left(\phi - 1\right) - 9.28\left(\phi - 1\right)^2\right] & 1.0 \le \phi_{50} \le 1.59 \\ 0.00218\,\phi^{14.2} & \phi_{50} < 1.0 \end{cases} \tag{10}$$

where:

$$W_i^* = \frac{\left(s = 1\right)gq_{b_i}}{u_*^3 F_i} \tag{11}$$

and F_i is the fraction of the bed surface sediment, calculated from a sample of the surface layer with sand and finer sizes (D<2 mm) excluded. Transport rates are calculated only for the gravel fraction of the surface layer; hence the subscripts s and g are used throughout the following explanation.

The parameter ϕ is formulated from a nested set of equations. The first of these is a hiding function:

$$\phi = \omega\phi_{sg}\left(\frac{D_i}{D_{sg}}\right)^{-0.0951} \tag{12}$$

The second equation is a function that accounts for changes in the mean grain size and sorting of the bed surface as the shear stress and transport rate increase. Parker (1990a) termed this a straining function:

$$\omega = 1 + \frac{\sigma_\phi}{\sigma_{\phi 0}}\left(\omega_0 - 1\right) \tag{13}$$

where σ_0 and ω_0 are functions of ϕ_{sg} given in figure 5. For typical values of sediment sorting and bed shear stresses not far above the threshold for bed load transport (say, 1.0<ϕ_{sg}<1.5), the function ω takes on values between 1.0 and about 0.8.

The third function is the relation for transport stage, expressed in terms of the Shields stress:

$$\phi_{sg} = \frac{\tau^*_{sg}}{\tau^*_{r_{sg}}} \tag{14}$$

where $\tau^*_{r_{sg}}$ is the reference Shields stress, assumed to be 0.0386 (Parker 1990a). The surface-based Shield stress τ^*_{sg} is defined as:

$$\tau^*_{sg} = \frac{u_*^2}{(s-1)gD_{sg}} \tag{15}$$

where u_* is calculated with the resistance relation presented earlier, Eq. 4, assuming $k_s = 2D_{90}$. Bed-load transport calculations are restricted to the main channel. For each flow of interest, the model calculates values of W^* for each size fraction and weights those values by the proportion of that size fraction on the bed surface, F_i. The instantaneous width-integrated bed load transport rate for each size fraction is then:

$$Q_{b_i} = \frac{W_i^* F_i B\, u_*^3 \rho_s}{(s-1)g} \tag{16}$$

where B is the channel width and u_* is calculated with respect to the main channel only. The predicted values of Q_{b_i} are summed over all size fractions to get the total bed load, Q_b.

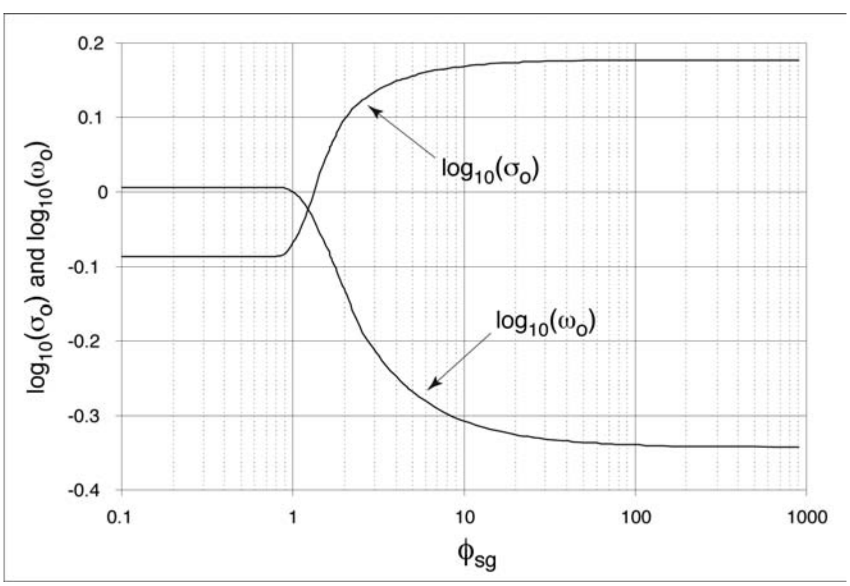


Figure 5. Parameters σ_0 and ω_o as functions of ϕ_{sg} in Parker's surface-based equation.

Substrate-Based Bed Load Equation of Parker-Klingeman-Mclean (1982)

The Parker-Klingeman-McLean equation (Parker and others 1982) computes transport rates on the basis of a single grain size—the median grain size of the substrate, $D_{50_{sub}}$. This equation likewise has three components:

$$W^* = \begin{cases} 11.2\left(1 - \dfrac{0.822}{\phi_{50}}\right)^{4.5} & \phi_{50} > 1.65 \\ 0.0025 \exp\left[14.2\left(\phi_{50} - 1\right) - 9.28\left(\phi_{50} - 1\right)^2\right] & 0.95 \le \phi_{50} \le 1.65 \\ 0.0025\, \phi_{50}^{14.2} & \phi_{50} < 0.95 \end{cases} \qquad (17)$$

where ϕ_{50} is the normalized Shields stress, formulated using the D_{50} of the substrate:

$$\phi_{50} = \frac{\tau_{50}^*}{\tau_{r_{50}}^*}\ , \qquad \tau_{50}^* = \frac{u_*^2}{\left(s-1\right) g D_{50_{sub}}} \qquad (18a,b)$$

and $\tau_{r_{50}}^*$ is a reference Shields stress with a value of 0.0876.

Here, the function W^* is modified slightly for ϕ_{50}<0.95 according to the surface-based equation of Parker (1990). The original equation sets $W^* = 0$ for ϕ_{50}<0.95. The modification in (17) does not change the equation substantially but it has the advantage of giving positive bed load transport rates for all flow conditions.

Parker and others (1982) did not specify an appropriate flow resistance relation for use with this equation. However, the equation is substrate based, thus k_s should be calculated with respect to the substrate. Based on limited data in Parker and others (1982), the ratio of surface D_{90} to substrate D_{50} ranges between 4.3 and 5.1 with an average value of ~5.35. Using this result and the previous assumption that the roughness height is twice of the surface D_{90}, we assumed:

$$k_s = 10.7 D_{50_{sub}} \qquad (19)$$

As with the previous equation, transport calculations are restricted to the main channel. Values of W^* are computed with Eq. 17 and the instantaneous transport rates are averaged over the width of the channel to get the total bed load:

$$Q_b = \frac{W^* B\, u_*^3 \rho_s}{\left(s-1\right) g} \qquad (20)$$

Substrate-Based Equation of Parker-Klingeman (Parker and Klingeman 1982)

The substrate-based bed load transport equation of Parker and Klingeman (1982) can be written as follows:

$$\frac{P_i\left(s-1\right) g Q_b}{f_i B u_*^3} = 11.2\left[1 - 0.853 \frac{\tau_{r_{50}}^*}{\tau_{50}^*}\left(\frac{D_i}{D_{50}}\right)^{\beta}\right]^{4.5} \qquad (21)$$

where p_i is the proportion of the bed load in the i-th size class; f_i is the fraction of the substrate in the i-th size class; τ^*_{50} is the Shields stress, formulated in terms of the median grain size of the substrate, $D_{50_{sub}}$; $\tau^*_{r_{50}}$ is the reference Shields stress, also formulated in terms of $D_{50_{sub}}$; D_i is the mean grain size of the i-th size class; and β is a hiding coefficient. The hiding coefficient, β, and reference Shields stress, $\tau^*_{r_{50}}$, are given by Parker and Klingeman as:

$$\beta = 0.018 \ , \qquad \tau^*_{r_{50}} = 0.0876 \tag{22a,b}$$

The same approach is used in calculating flow resistance as in the previous equation, with the assumption $k_s = 10.7\ D_{50_{sub}}$. Instantaneous transport rates are computed for each size fraction with (21) and averaged over the channel width:

$$Q_{b_i} = \frac{W_i^* \ f_i B \ u_*^3 \rho_s}{p_i\left(s-1\right)g} \tag{23}$$

The calculated values of Q_{b_i} are then summed over all size fractions to get the total bed load, Q_b.

Surface-Based Two-Fraction Equation of Wilcock (2001)

The surface-based two-fraction equation of Wilcock (2001) is a calibrated procedure that separates the bed load into two fractions—sand and gravel—and determines the reference shear stress for each fraction on the basis of bed load measurements. In theory, any suitable bed load transport equation can be calibrated using this approach. Wilcock (2001) recommended the following equations for estimating gravel and sand transport rates separately:

$$W_g^* = \begin{cases} 11.2\left(1 - 0.846\dfrac{\tau_{r_g}}{\tau}\right)^{4.5} , & \tau > \tau_{r_g} \\ 0.0025\left(\dfrac{\tau}{\tau_{r_g}}\right)^{14.2} , & \tau \le \tau_{r_g} \end{cases} \tag{24a}$$

and

$$W_s^* = 11.2\left(1 - 0.846\sqrt{\frac{\tau_{r_s}}{\tau}}\right)^{4.5} \tag{24b}$$

In these equations, the subscripts g and s refer to the gravel and sand fractions, respectively; τ_{r_g} is the reference shear stress for gravel; and τ_{r_s} is the reference shear stress for sand. Values for τ_{r_g} and τ_{r_s} are obtained from least squares regression based on bed load measurements. The separate values of W^* are then weighted by the respective fractions of gravel and sand on the bed, f_g, and $f_s = 1 - f_g$. Values of f_g and f_s are determined from a representative sample of the bed surface.

The relation used to estimate flow resistance is similar to the one used in the previous equations, with a slight difference in the assumed roughness, $k_s = 2D_{65}$. Instantaneous transport rates for the sand and gravel fractions are calculated separately:

$$Q_{b_g} = \frac{W_g^* f_g \rho_s B u_*^3}{(s-1)g} \;, \qquad Q_{b_s} = \frac{W_s^* f_s \rho_s B u_*^3}{(s-1)g} \tag{25a,b}$$

and then summed to give the total load,

$$Q_b = Q_{b_g} + Q_{b_s} \tag{26}$$

Surface-Based Relation of Wilcock and Crowe (2003)

Wilcock and Crowe (2003) developed a transport relation based on the full grain size distribution of the bed surface, including the sand. This relation includes an additional function that accounts for the nonlinear effect of sand content on gravel transport rates. The basic form of the equation is as follows:

$$W_i^* = \begin{cases} 0.002\phi^{7.5} & \phi < 1.35 \\ 14\left(1 - \dfrac{0.894}{\phi^{0.5}}\right)^{4.5} & \phi \geq 1.35 \end{cases} \tag{27}$$

where

$$\phi = \frac{\tau}{\tau_{r_i}}, \qquad \tau_{r_i} = \tau_{r_{50}} \left(\frac{D_i}{D_{50}}\right)^b \tag{28a,b}$$

The exponent in the hiding function b is calculated from:

$$b = \frac{0.67}{1 + \exp\left(1.5 - \dfrac{D_i}{D_m}\right)} \tag{29}$$

where D_m is the mean grain size of the bed surface. The reference shear stress for D_m is found using the Shields stress relation:

$$\tau_{r_m}^* = \frac{\tau_{r_m}}{(s-1)gD_m} \tag{30}$$

and an empirical function that accounts for the variation in τ_r^* with changes in sand content:

$$\tau_{r_m}^* = 0.021 + 0.015\exp(-20F_s) \tag{31}$$

where F_s is the percent of sand on the bed surface.

Values of W^* are calculated for each size fraction then weighted by the proportion of that size fraction on the bed surface, F_i. Those values are summed over all sizes to get the instantaneous width-integrated bed load transport rate:

$$Q_{b_i} = \frac{W_i^* F_i B u_*^3 \rho_s}{(s-1)g} \tag{32}$$

Procedure of Bakke and others (1999)

The procedure of Bakke and others (1999) uses site-specific measurements of bed load and bed material to calibrate two parameters in the equation of Parker and Klingeman (1982). The calibration procedure requires at least one, and preferably five to 10 bed load samples, plus samples of the bed material (surface or substrate). The procedure computes the hiding function exponent, β, and reference Shields stress for D_{50} (τ^*_{r50}), using a least squares fit to the calibration data. The procedure seeks to minimize the sum of squared differences between computed and measured transport rates based on different values of β and τ^*_{r50}. Transport rates are computed for each size fraction (surface or the substrate, depending on available data) and then summed to get the total bed load.

Data Input Requirements

Table 1 provides a list of the information and input variables needed to run the individual transport models. The first three rows list information common to all the models and include measurements of the reference-reach cross section, reach-average slope, and water discharge. Subsequent rows list particular sediment parameters derived from representative samples of the bed surface sediment, substrate, and/or bed load. The rationale for using surface versus substrate as input to a bed load transport model is discussed in Wilcock and others (2009). The models developed by Parker (1990a) and Wilcock and Crowe (2003) require data obtained from surface samples, otherwise known as pebble counts (table 1). The models developed by Parker and others (1982) and Parker and Klingeman (1982) require information obtained from bulk samples of the substrate (table 1). The approaches described by Wilcock (2001) and Bakke and others (1999) require actual bed load measurements, which serve as the basis for model calibration.

Table 1. Summary of input variables and information needed to run individual transport models.

	Parker	PKM	PK	W	WC	B
Channel cross section	X	X	X	X	X	X
Reach-average water surface slope	X	X	X	X	X	X
Discharge	X	X	X	X	X	X
Bed surface grain size distribution	X				X	X
• f_s, f_g				X		
Substrate grain size distribution			X			X
• D_{50}		X				
Bed load sample data				X		X

Software Operation

Model calculations and decisions are carried out in Visual Basic, a programming application that is included with recent versions of Microsoft Excel. This structure allows for straightforward cut-and-paste transfer of input and output data from one spreadsheet to another or any other program that accepts data in tabular format. Data must be entered in metric units. The program is designed to operate on WINDOWS-based PCs using Microsoft Excel, vers. 2000 and higher. The program does not run on Macintosh- or Unix-based platforms. Users should have a basic familiarity with the Microsoft Windows operating system and Microsoft Excel.

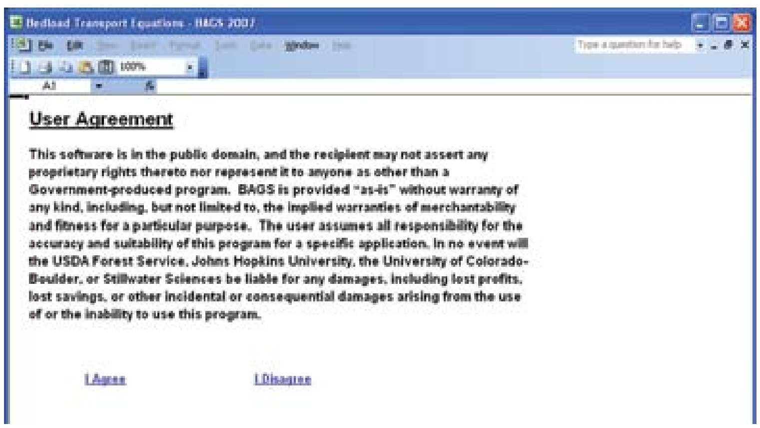
If you don't have the program for calculating bed load transport rates (BAGS), you may download a copy from the STREAM team website:

http://www.stream.fs.fed.us/publications/software.html.

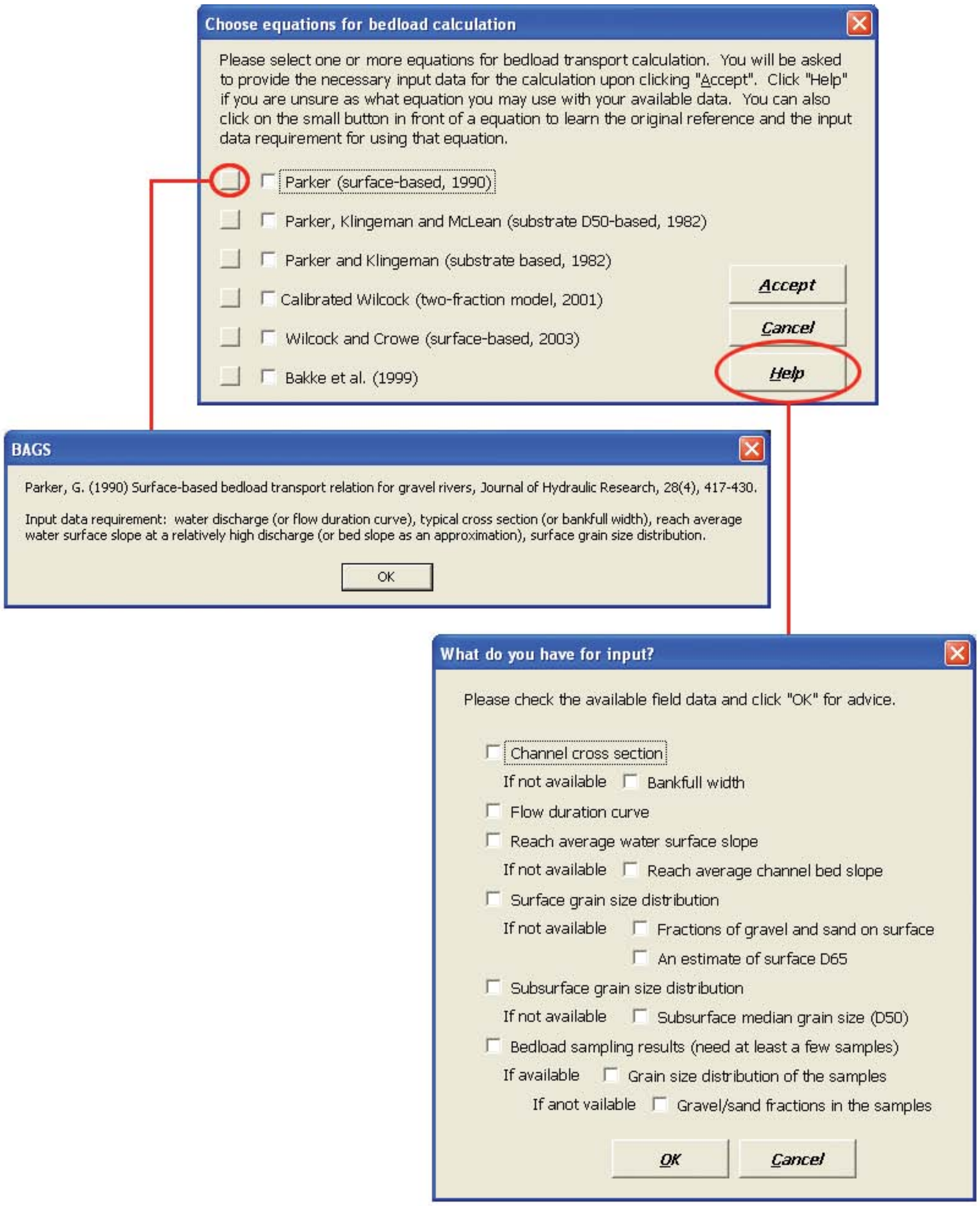
We recommend that you create a folder specific to your particular project and place the application program in this folder. Ancillary files or spreadsheets containing pertinent field data, including cross sections, water-surface profiles, discharge values, and/or sediment size distributions, should also be placed in the project folder. The steps involved in operating the software are listed below. In some cases, we have added sidebars that provide a comment or an explanation of the rationale behind an individual step or model calculation.

1. Before starting the program you should check the macro security level for your version of MS-Excel. It should be set at "Low" to run the BAGS program. To check the macro security level, start MS-Excel. Use the "Tools" menu and select "Options." Select the "Security" tab and click "Macro Security." Under the "Security Level" tab, enable "Low" and select "OK."
2. Start the program by clicking the BAGS icon. You will see a dialog box indicating that the program contains "macros." Select "Enable Macros."

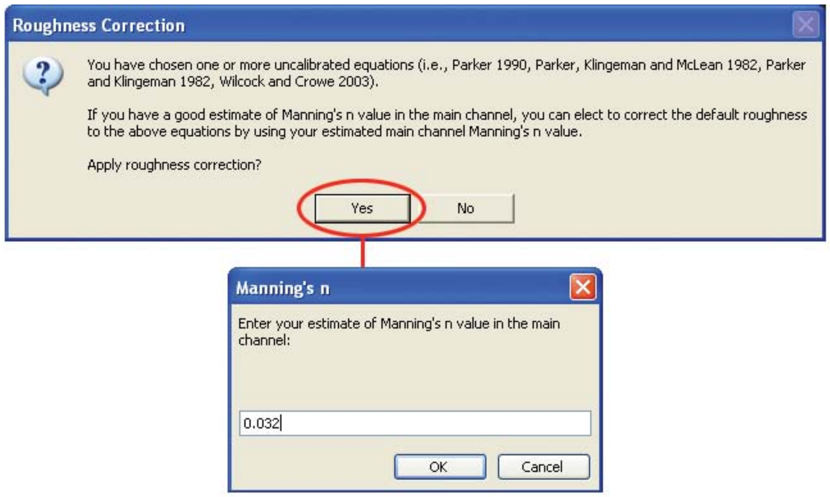
3. The BAGS background page will appear. Click the link to *apply bed load transport equations*. A users agreement will appear. Please read the conditions specified in the users agreement. If you agree with the terms, select "I Agree."



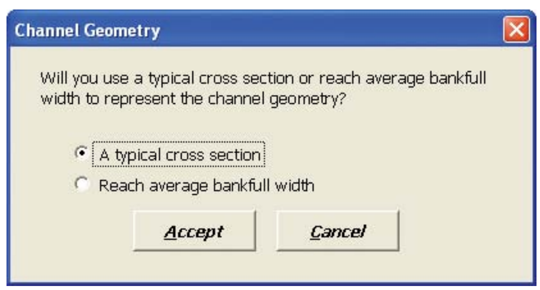
4. A box listing the equations will appear. If you are uncertain about which equation to use, select "HELP." If you would like additional information on an equation, including the original journal reference and a list of data input requirements, select the small box on the far left adjacent to the open box next to the equation. When you are ready to proceed, check the open box next to the equation and select "ACCEPT."



5. You will now see a dialog box asking if you would like to apply a roughness correction. If you have a field-based estimate of Manning’s *n*, select “YES” and enter that value when the next dialog box appears. If you don’t have an estimate of Manning’s *n* but would like to apply a roughness correction anyway, we suggest you consult the references listed earlier in the section on flow resistance.

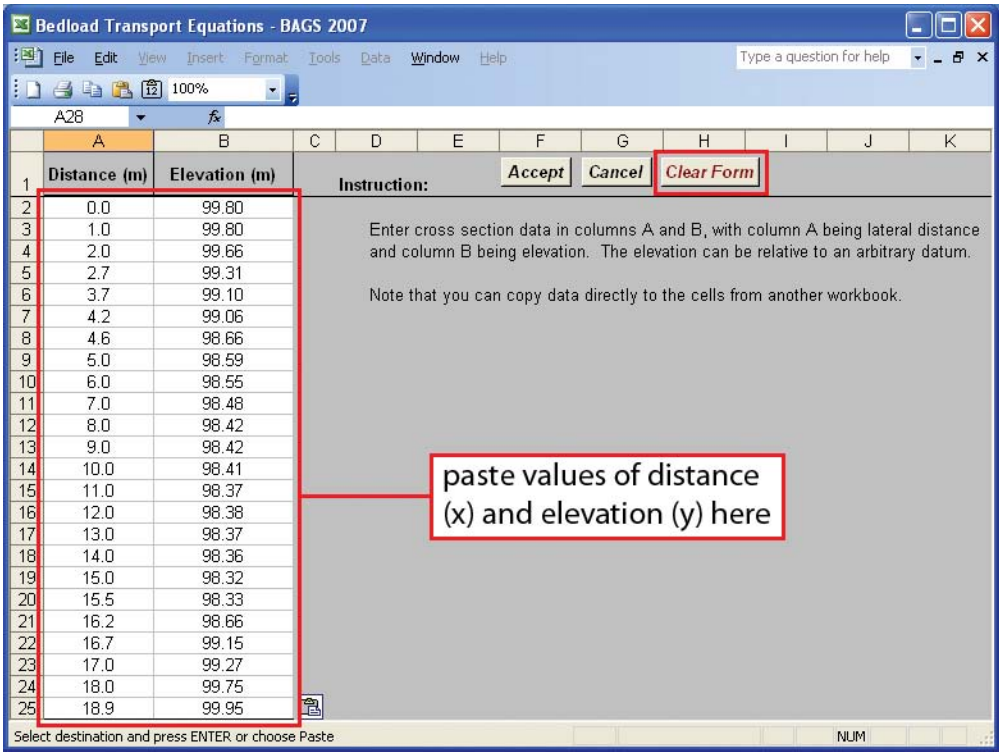


6. A dialog box will appear indicating two choices for channel geometry:

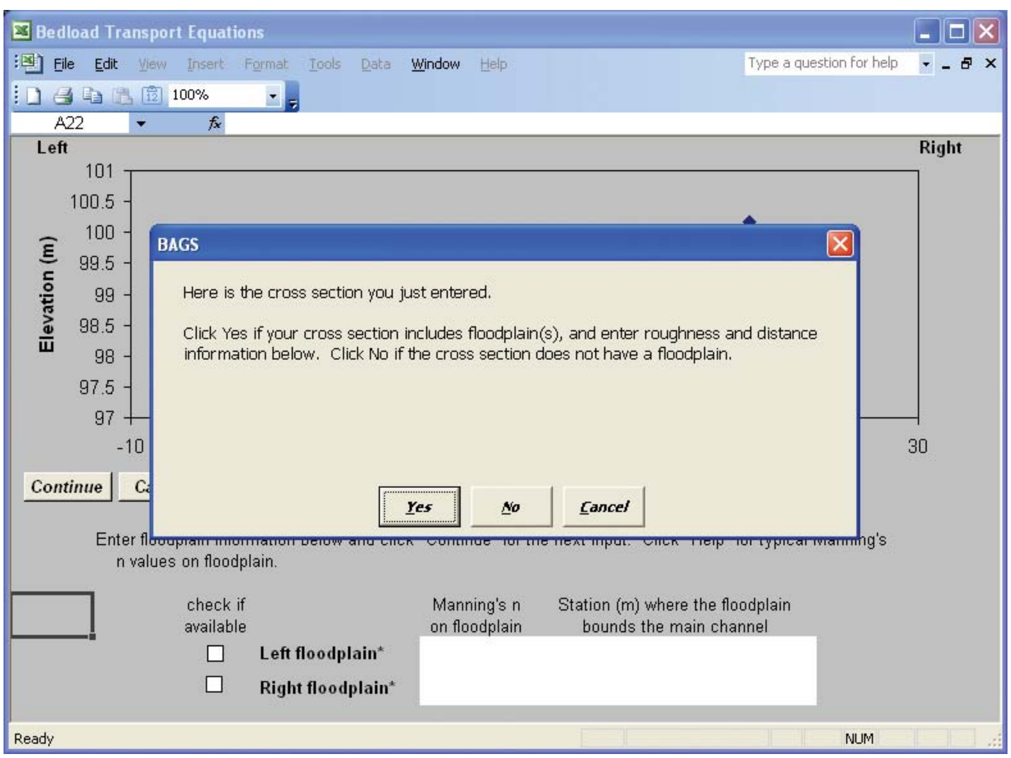


Ideally, you should have cross section data available to enter at the next step; however, the software can approximate flow conditions for a known width and discharge by assuming a rectangular cross section.

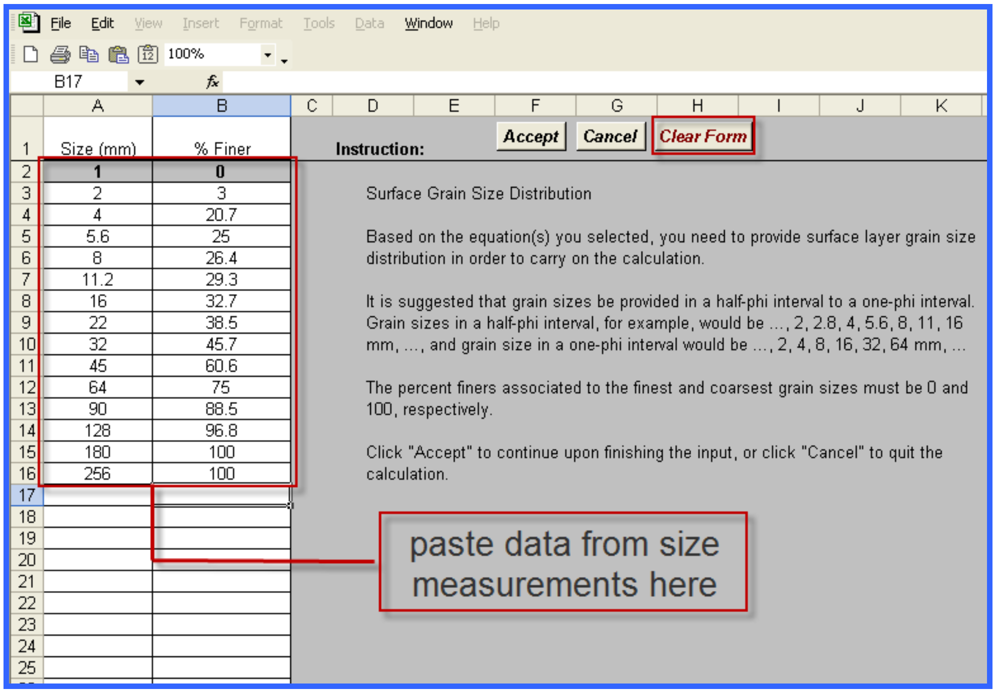
7. The next box allows you to enter data for the reference reach cross section. Select “CLEAR FORM” and enter values of distance and elevation for the cross section.



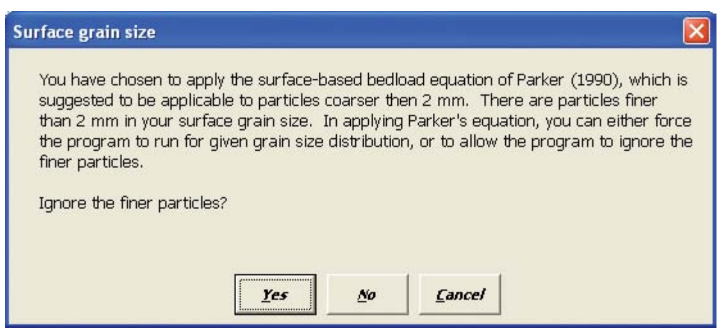
8. A new page will appear with a plot of the reference reach cross section. If you want to include the floodplain in the hydraulic calculations, you must assign values of Manning’s *n* to the left- and right-portions of the floodplain.



9. A new page will appear, asking for input on the grain size distribution. Any specific intervals for the grain size, *D*, may be used, however, the sizes must be listed from lowest to highest, for example: 1, 2, …., 128 mm. Select "CLEAR FORM" and enter the percent finer for each size class in the appropriate columns.



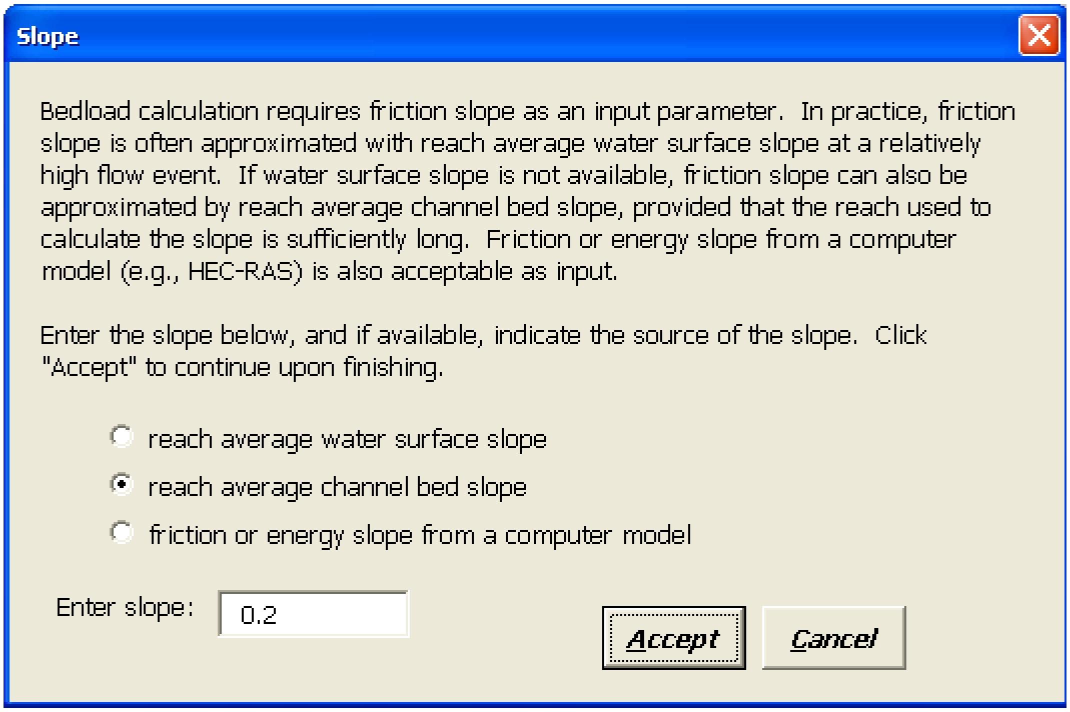
Note: Depending on the bed load equation you have selected, you may then see another dialog box asking if you would like to include grain sizes less than 2 mm in the calculation.



Sidebar 1

Comment: Parker (1990a) formulated his surface-based relation by excluding sand from the analysis and computation. Subsequently, Wilcock and Crowe (2003) showed that sand has a strong influence on transport thresholds and transport rates. Other work indicates that, in many rivers, the bed load is predominantly sand (Hassan and Church 2001; Lisle 1995; Mueller and others 2005), consistent with field data showing that the substrate is 20 to 30 percent sand (Pitlick and others 2008). Given this information, we don't see a clear reason for excluding sand from the calculation unless it can be determined that these sizes will move in suspension.

10. The next dialog box will ask you to select one of potentially three values for slope.



Sidebar 2

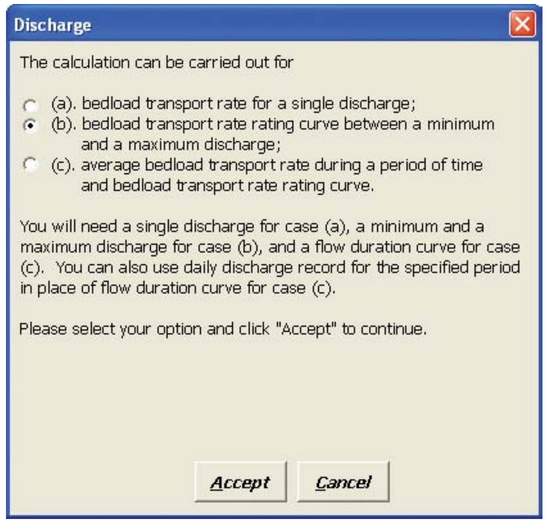
Note: If you select option (b), you will need to supply estimates of minimum and maximum discharges, otherwise the program may stall or give unreasonable results. If you are not sure what to enter for minimum and maximum values, we can offer several suggestions:

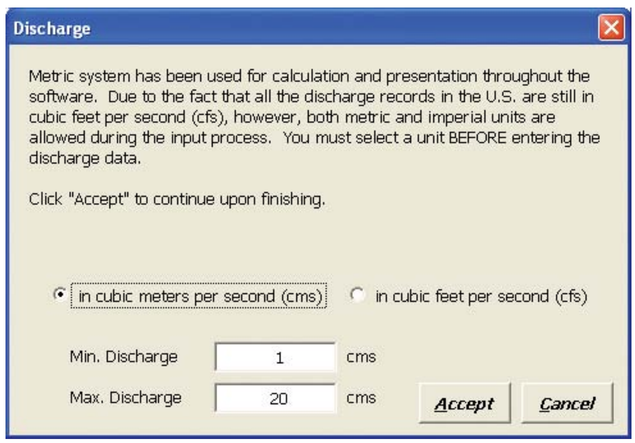
1. If the site is gaged and the gage record includes at least 10 years of peak-flow values, you could use the lowest and highest values to define a reasonable range of discharges or

2. If the site is not gaged, but you have a rough idea of the bankfull discharge (or perhaps an estimate of an average-size flood, such as the 2-year flood), we suggest setting the minimum discharge to approximately half this value and setting the maximum discharge to 2 (or maybe 4) times this value.

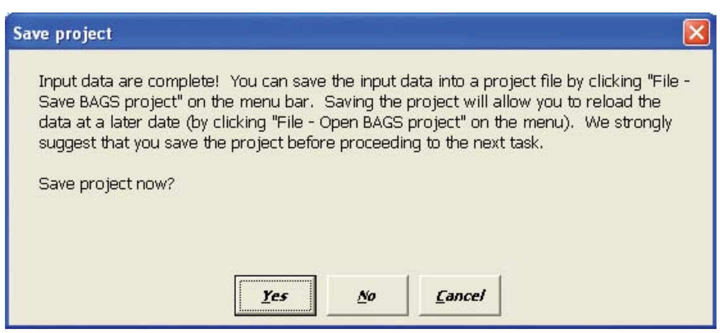
These limits are somewhat arbitrary, but likely to encompass most of the range over which bed load transport occurs (this point is discussed in more detail in the Assessing Model Output section).

11. The next dialog box asks for information on discharge.

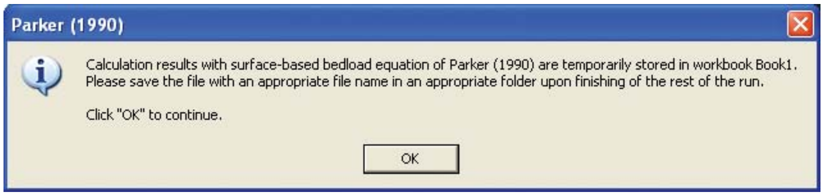




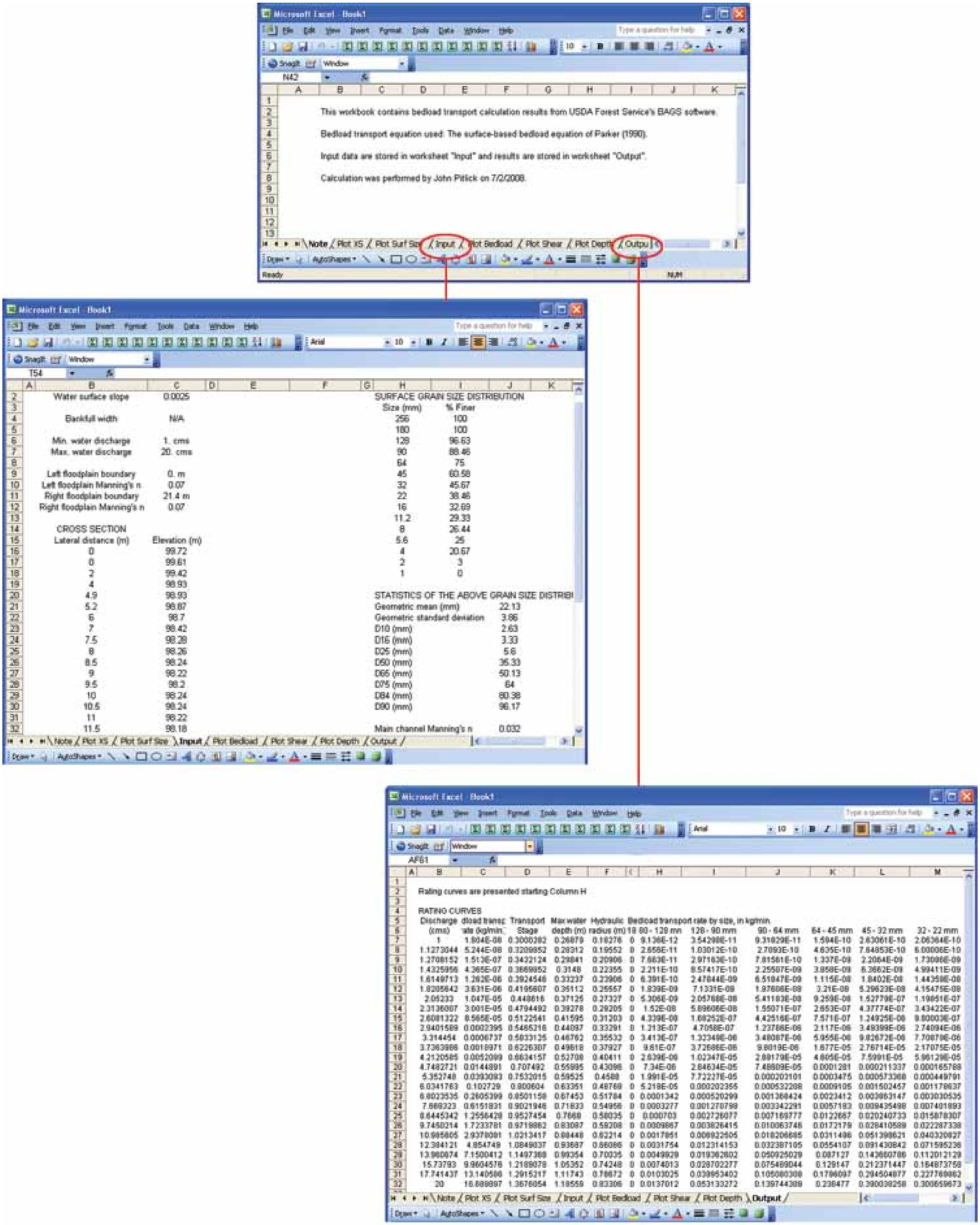
12. At this point, the steps for entering data are complete. The next dialog box will ask if you would like to create a file to save the input data.



13. The program performs a series of calculations and stores the results in a new workbook. When the calculations are complete the following box will appear.



14. The new workbook contains eight separate tabs. The first tab lists the equation(s) used, locations of the input/output data, and user and date. Other tabs list input/output values or show plots generated for the particular model run. You should check the Input tab to ensure that the parameter values and data were entered correctly. The Output tab lists values of discharge (m^3/s), bed load transport rate (kg/min), transport stage, τ / τ_r (dimensionless), maximum water depth, hydraulic radius, and, depending on the equation specified, bed load transport rates of individual size fractions.


This workbook contains bedload transport calculation results from USDA Forest Service's BAGS software.
Bedload transport equation used: The surface-based bedload equation of Parker (1990).
Input data are stored in worksheet "Input" and results are stored in worksheet "Output".
Calculation was performed by John Pitlick on 7/2/2008.
Water surface slope
Bankfull width
Min. water discharge
Max. water discharge
Left floodplain boundary
Left floodplain Manning's n
Right floodplain boundary
Right floodplain Manning's n
CROSS SECTION
Lateral distance (m)
Elevation (m)
SURFACE GRAIN SIZE DISTRIBUTION
Size (mm)
% Finer
Geometric mean (mm)
Geometric standard deviation
Main channel Manning's n
Rating curves are presented starting Column H
RATING CURVES

Assessing Model Output

Given that bed load transport rates can vary by many orders of magnitude, almost any result produced by the BAGS model can be considered "reasonable." The questions are: What do we mean by reasonable? How do we distinguish between a potentially valid result and one that is implausible, and how do we assess model output if there are no data against which to compare the results? The sections below discuss possible strategies for evaluating the reasonableness of model results, starting with some comments about the data itself and potential input errors. The subsequent section goes through several examples illustrating how and why model results can differ from field measurements. The final section focuses on the common situation where no measurements of bed load transport have been made, thus there is little basis for comparison.

How Good Are the Raw Data?

Before discussing approaches for interpreting model output, it is worth reminding ourselves that a model calculation is only as good as the input data. The phrase "garbage in, garbage out" certainly applies in this case.

All bed load transport relations are sensitive to estimates of the available grain sizes and the available shear stress. Small differences in the estimation of these two values can lead to very large (orders-of-magnitude) differences in calculated transport rates. In addition to potential errors and uncertainties in the input variables, there may be conditions within a watershed that severely limit the usefulness of bed load transport calculations. The following points are important to consider:

- Is the site or the watershed appropriate for this type of analysis? Bed load transport relations are formulated with two assumptions in mind: (1) the mass flux of sediment is related in a consistent way to the physical properties of the flow (force or power per unit bed area) and (2) all the grain sizes capable of being transported by a particular flow are indeed available to be transported. Bed load transport equations predict the bed load transport capacity, which Gilbert (1914, p. 35) defined as the "maximum load a river can carry," presumably for a given flow and sediment size. Generally, this definition only applies to alluvial channels where the bed and banks are made of sediment that was carried by the river itself, as opposed to some other geomorphic process (for example, mass-wasting or glacial processes). This can be a serious limitation in actively eroding bedrock channels and mountain streams where the sediment supply is driven more by the rate of weathering and/or hillslope transport than by fluvial processes. In these cases, the channel potentially transports only as much sediment as is supplied, and the sediment flux bears only a weak relation to the flow strength. Transport under these conditions is said to be "supply limited," and there is no reason to expect that a transport relation will predict the sediment flux accurately.

- Is the bed material very heterogeneous? Can you determine, at least qualitatively, whether the bed material samples are representative of the reach of interest? The bed sediment in gravel channels can vary significantly from place to place, thus a surface or substrate sample from only one location may not yield a grain size distribution that is representative of that setting. Figure 6 shows grain size distributions of the surface and substrate sediment sampled across a single meander bend within a 130-m reach of the Colorado River in Rocky Mountain National Park, Colorado. This is a relatively stable gravel-bed river with a mildly sinuous channel pattern, yet these measurements show that there is a wide range in grain size of both the surface and the substrate. The variation in grain size is likely to be much higher in morphologically complex channels (braided or wandering rivers), thus sampling intensity should be increased to determine particle size distributions more accurately. The clearest guidance on procedures for sampling and analysis of the bed sediment in gravel channels can be found in Church and others (1987) and Bunte and Abt (2001). We cannot stress the importance of taking as much time as necessary to obtain representative samples of the bed material—it makes no sense to spend hundreds of hours taking measurements of water discharge or channel properties and then spend 1 hour sampling the bed material. The problems associated with a potentially inaccurate estimate of the grain size are illustrated later in the discussion of transport estimates for the South Fork Samon River (p. 35).

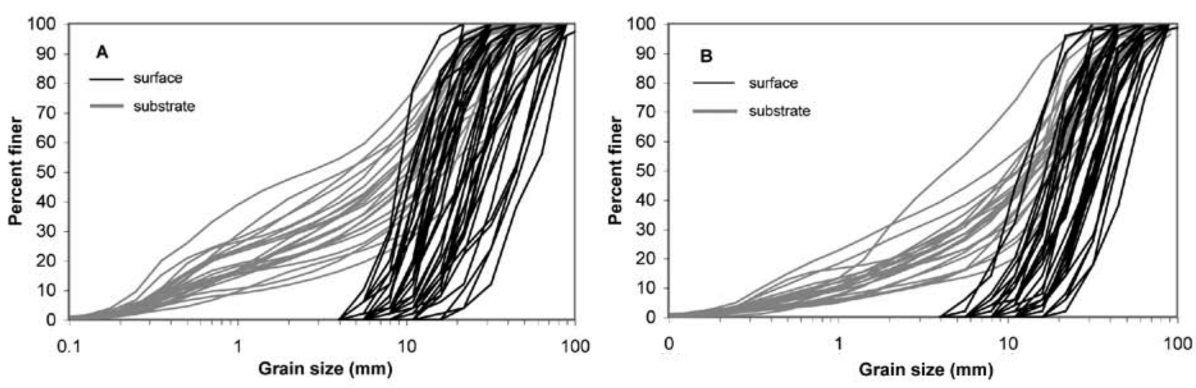


Figure 6. Grain size distributions of the bed surface and substrate of the Colorado River in Rocky Mountain National Park, CO. The samples were taken across a single meander bend, at different locations corresponding to (A) the middle portion of bend and (B) the outer portion of the bend (source: Clayton and Pitlick 2007).

- Are the estimates of boundary shear stress reasonable? The variations in shear stress within a channel reach are likely to be at least as large as the variations in grain size. The problems of estimating shear stress are particularly important in high-gradient channels (slopes greater that ~1 percent) where a significant proportion of the total shear stress acting on the streambed may be “lost” due to form drag on immobile boulders and logs (Buffington and Montgomery 1999; Mueller and others 2005; Wiberg and Smith 1990; Wilcox and Wohl 2006). The BAGS model attempts to correct for these effects, but the problem is not restricted to small, headwater streams. Natural undulations in bed topography caused by pools and riffles force changes in depth and velocity, thus altering the distribution of shear stress (Sear 1996; Whiting and Dietrich 1991). The BAGS model is not likely to yield accurate estimates of bed load transport in highly sinuous channels, braided channels, or channels with sharp changes in gradient.

Comparisons of Model Output With Bed Load Measurements

As noted in Wilcock and others (2009), a few measurements of bed load can aid significantly in assessing the uncertainty of a transport calculation. However, even in a best-case scenario, one should expect that bed load measurements taken at the same location at the same discharge will vary by at least an order of magnitude. This is especially true of samples taken at flows near the threshold for motion. The following examples illustrate the range of results one can expect in comparing output from the BAGS model with field measurements of bed load. The data sets used for these illustrations are based on measurements taken in different types of gravel-bed streams and rivers in the western United States. The discussion emphasizes “goodness-of-fit,” not as a statistical concept, but as a means of comparing observed values with predicted values.

Visualizing Bed Load Transport as a Function of Discharge

Measurements of bed load taken in the field and laboratory indicate that transport rates increase by orders of magnitude for relatively small changes in discharge and shear stress. The nonlinearity in transport processes is reflected by the high value of the exponent in the relation between bed load and shear stress. In the equation of Wilcock and Crowe, for example, $W^* \propto \tau^{7.5}$ in the range of low-moderate shear stresses, $\tau < 1.35\tau_r$. Flows in this range typically carry a high percentage of the total annual bed load (Andrews 1994; Andrews and Nankervis 1995; Emmett and Wolman 2001; Mueller and others 2005; Torizzo and Pitlick 2004; Van Steeter and Pitlick 1998; Whiting and others 1999). For these reasons we recommend plotting the relation between flow and transport using logarithmic scales for both axes. Otherwise, it is essentially impossible to visually assess the quality of the data, especially in the range of flows where most of the transport occurs.

Field hydrologists are accustomed to using water discharge, rather than shear stress, as the primary index of flow properties. Discharge is commonly measured in the field, and values of discharge associated with individual transport measurements are typically listed in published reports, whereas estimates of shear stress are not. Discharge thus emerges as a natural variable for associating transport rates and flow.

A statistical goodness-of-fit-test helps in judging the strength of the relation between bed load discharge, Q_b, and water discharge, Q, but it may be more instructive to ask: Do the field data follow the expected (modeled) trend? To answer this question, we suggest using a simple test focusing on the parameters of a power-law relation, $Q_b = aQ^b$, where a is a coefficient and b is the slope of the observed transport relation. If the observed exponent b differs greatly from the expected value, then you might want to look more carefully at the measurements. To find the expected value of b, we combine the continuity equation with the Manning equation and write a relation for shear stress as a function of discharge:

$$\tau = \rho g \left(\frac{Qn}{B} \right)^{0.6} S^{0.7} \tag{33}$$

where the variables as defined in the Bed Load Transport equations section. This equation indicates that for constant values of ρ, g, n, B and S, τ varies with the 0.6 power of Q (we should note, however, that in typical channels, n will decrease with Q, and B will increase, thus the exponent may be expected to be less than 0.6—this depends on the particular site characteristics). Using this result, we can recast one of the transport relations discussed earlier in terms of Q. The Wilcock and Crowe equation for low-moderate transport stages ($\tau<1.35\tau_r$) is used as an example. This equation can be written as:

$$Q_b = \frac{0.002\rho_s}{(s-1)g} B \left(\frac{\tau}{\rho} \right)^{1.5} \left(\frac{\tau}{\tau_r} \right)^{7.5} = kB(\tau)^{9.0} \tag{34}$$

where Q_b is the total bed load (mass transport rate, integrated across the channel width, B) and k is a value incorporating the various constants, plus the value of τ_r. Collecting the various terms and writing the total bed load as a function of discharge we get:

$$Q_b \propto (Q)^{5.4} \tag{35}$$

The exponent of 5.4 in Eq. 35 is specific to this example, and we should not think of it as a hard number. We say this for several reasons. First, the specific value depends on the exponent in the transport relation (b = 7.5 in the relation used in this example). Second, the derivation of (35) was simplified by assuming B, n, and S were constant, which is not likely to be the case in natural channels. Third, as Barry and others (2004) have suggested, the rating curve exponent may be influenced by other factors, such as runoff regime (rainfall versus snowmelt) and sediment supply. Nevertheless, in the absence of strong constraints on bed load transport, or systematic errors in sampling, Eq. 35 serves as the basis for interpreting observed relations between Q and Q_b. This point is illustrated in figure 7, which shows relations between discharge and bed load transport in two small gravel-bed streams, Oak Creek and Halfmoon Creek. Bed load was measured in Oak Creek using a vortex sampler (Milhous 1973), whereas bed load was measured in Halfmoon Creek using a series of quasi-stationary traps (Bunte and Swingle 2005). These data are of high quality, and in both cases there is a strong correlation between bed load transport rate and discharge. The exponents are slightly different from each other, but they bracket the proposed value of 5.4 given in Eq. 35.

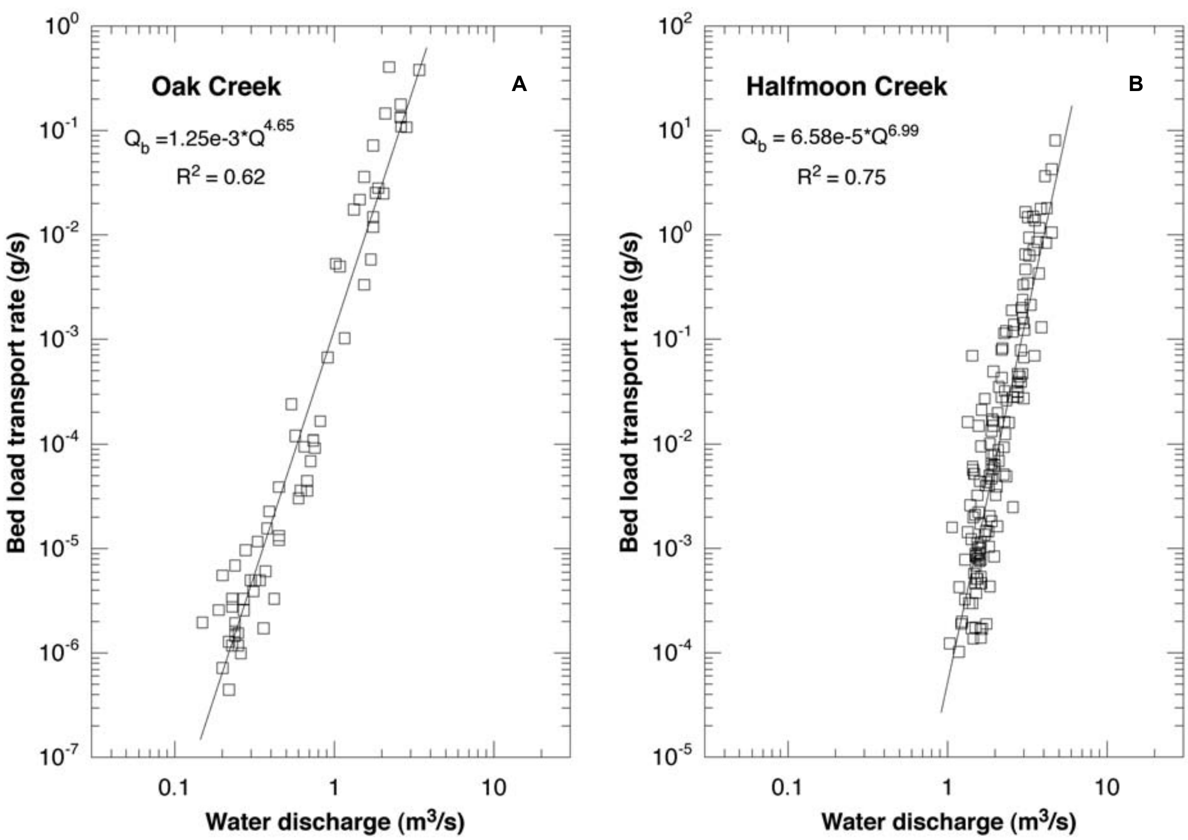


Figure 7. Observed relations between discharge and bed load transport. Data for Oak Creek from Milhous (1973); data for Halfmoon Creek from K. Bunte and Swingle 2005.

The steep slopes exhibited by these data reinforce the point made earlier that transport rates can vary by several orders of magnitude over the range of observed discharges. Small differences in discharge produce large differences in transport rate. More to the point, the data depicted above compare favorably with the derived transport relation (35), thus we might use these examples as a basis for comparing measurements in other settings where there is more uncertainty in the data. Four additional examples follow, with the predicted transport relations added for comparison. The predicted relations are developed using the surface-based equation of Parker (1990a) and appropriate data and information from field studies of flow and bed load transport in streams and rivers in Idaho (King and others 2004).

Selway River Near Lowell, ID

The Selway River is an example of a large self-formed gravel-bed river with no apparent limit to sediment supply. The data used in this example are based on measurements taken at a gauging station operated by the USGS, station number 13336500. Site characteristics are described as follows:

- Slope = 0.0021
- Surface D_{50} = 186 mm
- Subsurface D_{50} = 24 mm
- Drainage area = 4947 km^2
- Bankfull discharge = 652 m^3/s

Most of the bed load measurements at this location were taken during periods of peak snowmelt runoff in 1994 and 1995. Several additional samples were taken during floods in December 1995 and May 1997. The data set includes 72 paired measurements of discharge and bed load; field surveys of bed- and water-surface slope; and multiple samples of the surface and substrate. Bed load was measured with the Helley-Smith sampler, with 40 percent of the samples taken at discharges greater than one-half of the bankfull discharge.

The data from the Selway River form a very tight relation between water discharge, Q, and bed load discharge, Q_b (fig. 8A). A power law fit of these data yields the equation:

$$Q_b = 3.85\text{E-}12\ Q^{4.92} \qquad (36)$$

where Q is in cubic meters per second and Q_b is in metric tons per day. This relation is statistically significant, with $r^2 = 0.92$ and $p<0.001$. The exponent in this equation is relatively high (~5), similar to the best-fit relations for Oak Creek and Halfmoon Creek. The panel to the right (fig. 8B) compares the observed transport rates with the predicted transport rates estimated using the surface-based equation of Parker (1990). In this case, the predicted bed load transport rates match the observations quite closely. This result is encouraging, but somewhat of an exception—this is one of the few data sets we have worked with where the transport relation fits the data closely with no tuning or adjustment in the parameters.

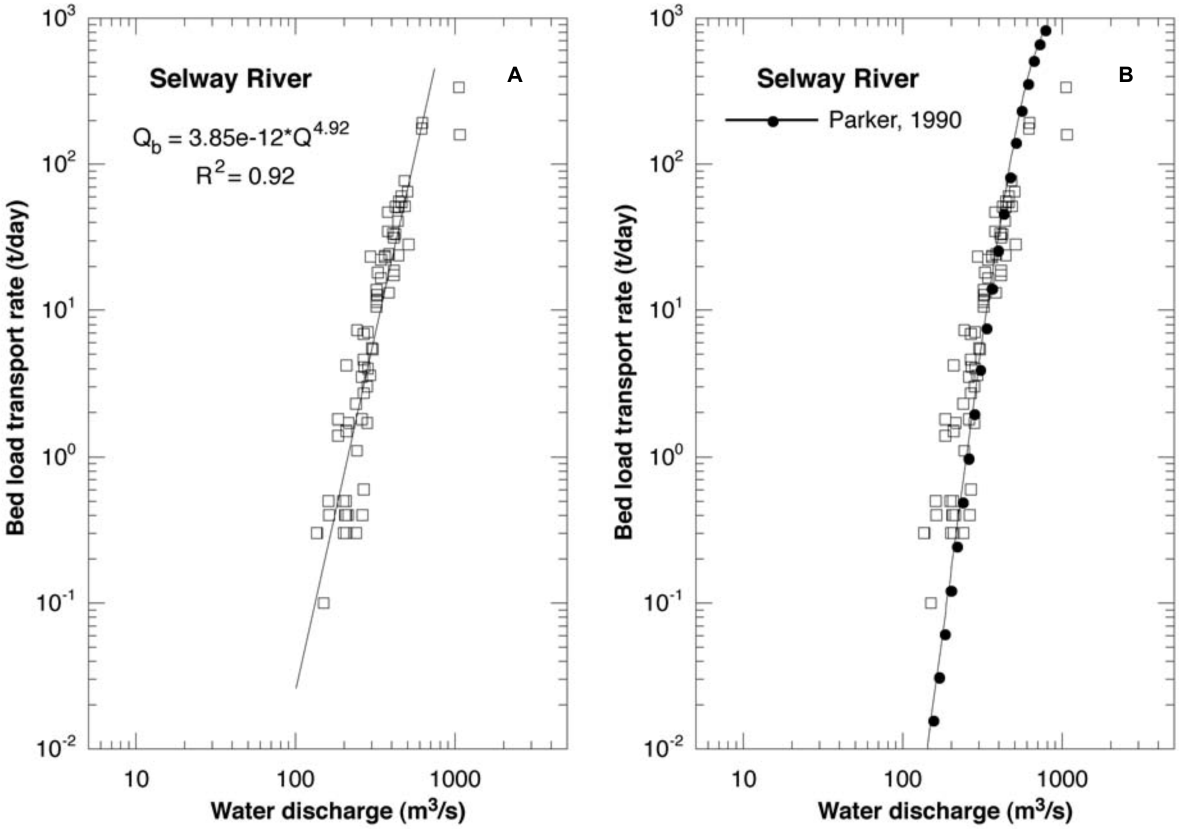


Figure 8. Relations between discharge and bed load transport rate, Selway River near Lowell, Idaho, USGS, station number 13336500. (A) Shows a least squares fit of the data and (B) shows the predicted relation obtained with the surface-based equation of Parker (1990a), with sand-size fractions included.

Rapid River

The Rapid River is an example of a steep gravel- and cobble-bed river. The site description indicates that the reach is bordered by a floodplain, thus it would be reasonable to conclude that, even if the reach is steep, the channel is self-formed. This gage is operated by the U.S. Forest Service. Site characteristics are:

- Slope = 0.0108
- Surface D_{50} = 79 mm
- Subsurface D_{50} = 16 mm
- Drainage area = 280 km^2
- Bankfull discharge = 17.7 m^3/s

Bed load samples were taken at this location from 1990 through 2004; the majority of samples were taken during spring runoff. The data set for this station includes: 190 paired measurements of discharge and bed load; surveys of bed- and water-surface profiles; and samples of the surface and substrate. Bed load was measured with the Helley-Smith sampler, with 38 percent of the bed load samples taken at discharges greater than one-half of the bankfull discharge.

The observed transport relation for the Rapid River is relatively strong (fig. 9), although the data exhibit considerable scatter for measurements taken below ~10 m^3/s. The question is: How much weight should we give the low-flow samples? Using all data, the best-fit equation is:

$$Q_b = 0.0086\ Q^{2.23} \tag{37a}$$

This relation is statistically significant ($r^2 = 0.59$; $p<0.001$). However, it is evident that at high discharges this equation would underestimate bed load transport rates by more than an order of magnitude. Thus, if we consider only flows greater than 10 m^3/s (~60 percent of bankfull), the relation formed by the data is much different. The least squares equation in that case is:

$$Q_b = 6.74\text{E-}06\ Q^{4.88} \tag{37b}$$

which is also statistically significant ($r^2 = 0.88$; $p<0.001$). This second relation provides a better fit to the high-flow values and the exponent is greater than 4, similar to the equations given in the previous examples. This example illustrates a common condition in gravel rivers where bed load transport rates at low discharges are influenced by differences in the availability of sand-sized sediment, which may be supplied from sources outside the channel. At discharges greater than 10 m^3/s, particles with the surface layer are beginning to move and the bed starts to become the primary source of sediment. Above this point, the modeled bed load transport relation (Parker 1990) matches the observations quite well.

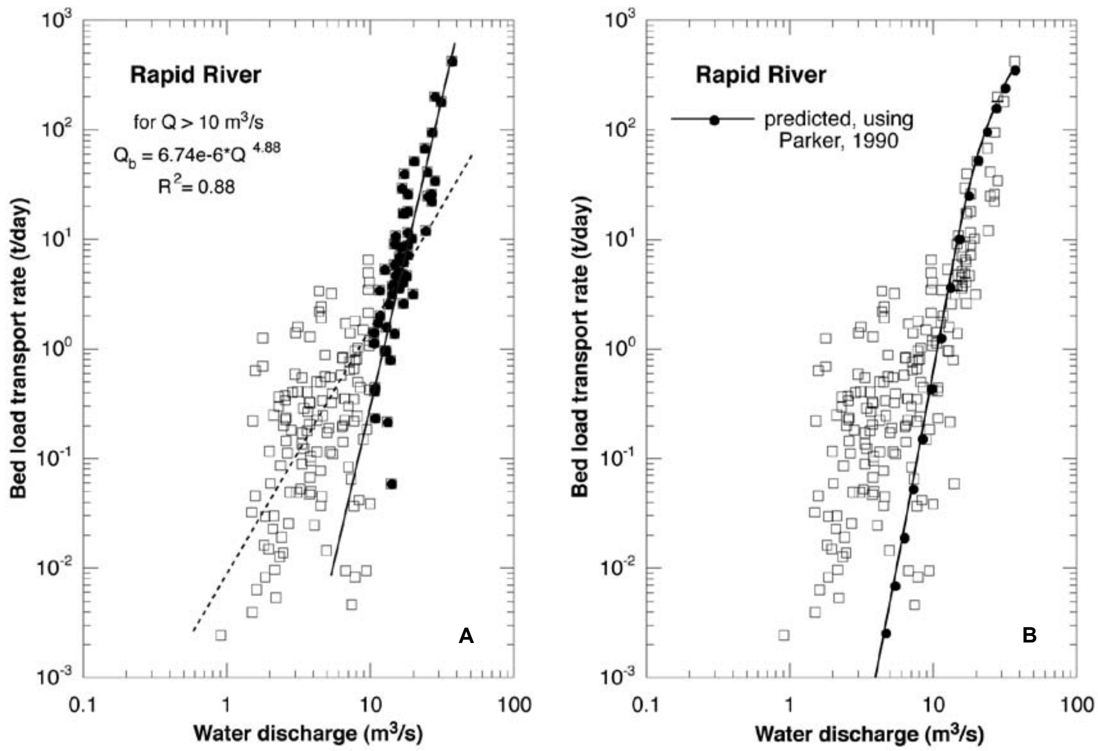


Figure 9. Relations between discharge and bed load transport rate, Rapid River, Idaho. (A) Shows two least squares relations, one for the complete data set (dotted line) and the other for Q>10m^3/s (solid line) and (B) shows the predicted transport relation obtained with the surface-based equation of Parker (1990a).

South Fork Salmon River Near Krassel Ranger Station, ID

The South Fork Salmon River is an example of a moderate-sized gravel river that carries a moderately high sediment load. The site description indicates that the study reach is bordered by a floodplain, thus it appears that the channel is self-formed. This station is operated by the USGS (station 13331070).

- Slope = 0.0025
- Surface D_{50} = 38 mm
- Subsurface D_{50} = n/a
- Drainage area = 855 km^2
- Bankfull discharge = 70.8 m^3/s

Bed load samples were taken at this location over two time periods, 1985 to 1986, and 1994 to 1995. The majority of samples were taken during spring runoff. The data set includes: 130 paired measurements of discharge and bed load; surveys of bed- and water-surface profiles; and samples of the bed surface (no substrate samples were taken at this site). Bed load was measured with the Helley-Smith sampler, with 57 percent of the samples taken at discharges greater than one-half the bankfull discharge.

The observed transport relation at this site appears to be relatively good (fig. 10A). The equation of the trend line is:

$$Q_b = 1.97\text{E-}04\ Q^{3.00} \qquad (38)$$

which is statistically significant (r^2 = 0.62 and p<0.001). The exponent in this equation is lower than in previous examples, and it appears that the fitted line has a lower slope than the trend formed by the data. This is not an uncommon result in fitting a power law to a data set that is not log-linear, or a data set that might include outliers. Eliminating the five values corresponding to Q<10 m^3/s gives an exponent of about 4, similar to the previous examples.

In this example, the comparison between observed and modeled transport rates produces an interesting result (fig. 10B). An initial calculation based on Parker's surface-based relation gives the solid curve to the right of the data. This curve parallels the data, but is offset to the right, meaning the predicted transport rates are much less than the observed transport rates. The difference arises because the reference Shields stress used in the calculations is either too high or the grain size measured in the field is coarser than the sediment being supplied to the river. A second calculation, based on the calibration approach of Wilcock (2001), gives the relation indicated by the dotted line. This relation fits the data well, but the difference between the two curves suggests either that the sampled grain size is not representative of the reach or that much of the bed load is derived from sources other than the channel itself.

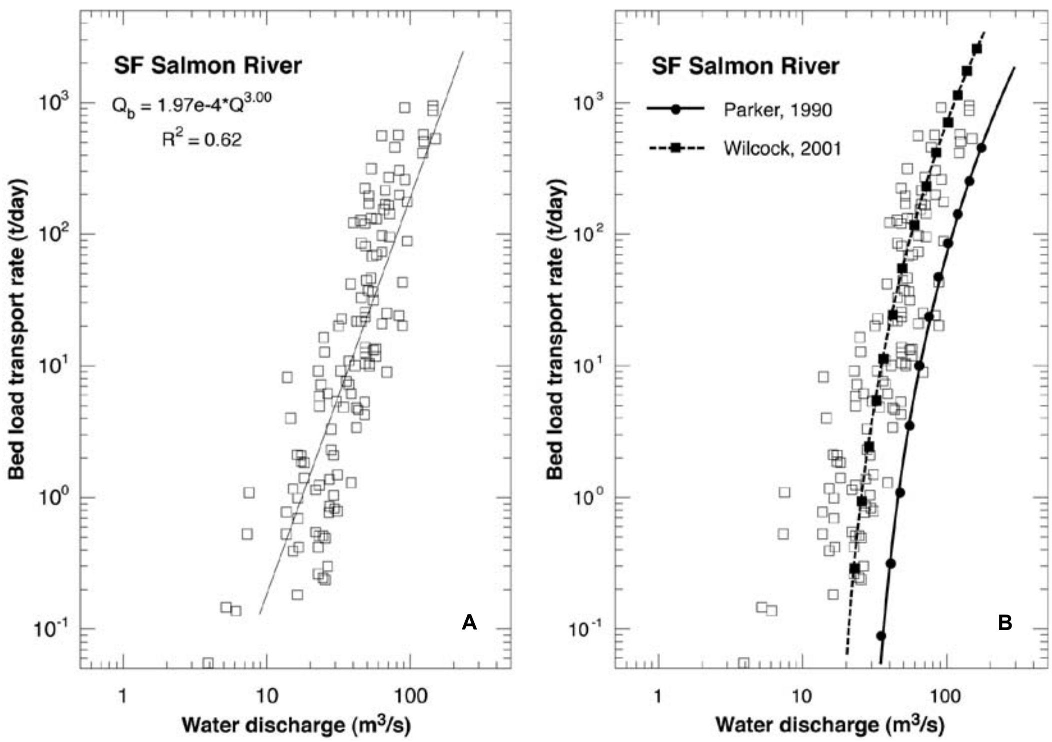


Figure 10. Relations between discharge and bed load transport rate, South Fork Salmon River, Idaho. (A) Shows a least squares fit of the data and (B) shows modeled transport relations. The solid line indicates the surface-based equation of Parker (1990a) and the dotted line indicates the calibration approach of Wilcock (2001).

Trapper Creek

Trapper Creek is an example of a small cobble-bed stream (drainage area = 20.7 km^2) in a forested area. The study reach is steep (~ 4 percent); however, the photograph of the site suggests that the channel is self-formed and possibly bordered by alluvial banks. The U.S. Forest Service operates this station.

- Slope = 0.0414
- Surface D_{50} = 85 mm
- Subsurface D_{50} = 17 mm
- Drainage area = 20.8 km^2
- Bankfull discharge = 2.6 m^3/s

Bed load samples were taken at this location from 1986 through 2001. Many of these samples were taken during spring runoff. The data set includes: 115 paired measurements of discharge and bed load; surveys of bed- and water-surface profiles; and samples of the surface and substrate. Bed load was measured with the Helley-Smith sampler. In contrast to the previous examples, only 17 percent of the samples were taken at discharges greater than one half the bankfull discharge.

The measurements at this site form a relatively well-defined transport relation, although there is significant scatter in the data (fig. 11A). A least squares fit of the full data set gives:

$$Q_b = 0.35\ Q^{1.80} \tag{39a}$$

This relation is statistically significant (r^2 = 0.60; p<0.001); however, the exponent in the equation and the slope of the line are somewhat low in comparison to the values given in the previous examples. If we restrict the regression to flows greater than 1.5 m^3/s, which is the point where we might expect clasts within the surface layer to start moving (60 to 70 percent of the bankfull discharge, see below), we get a relation with slightly steeper slope and a higher exponent:

$$Q_b = 0.124\ Q^{3.32} \tag{39b}$$

This relation is not as strong as the previous relation but still statistically significant (r^2 = 0.31; p = 0.0014). Given this result, we might assume that a transport equation would fit the high-flow data reasonably well, as we saw in the example of the Rapid River. However, when we plot the modeled relation, we see that it lies far to the left and above most of the observations (fig. 11B), suggesting that predicted loads are two to three orders-of-magnitude higher than observed loads. This level of uncertainty would be unacceptable in most situations. It is even less reassuring to know that this type of discrepancy is relatively common, particularly in steep channels (and, it is exactly this type of problem that leads people to question the utility of bed load transport equations). However, there is a straightforward explanation in this case, which probably holds true for many other high gradient channels. Specifically, Trapper Creek is a steep stream (S = 0.04) with very coarse bed material (D_{90} = 136 mm), thus it is likely that a significant proportion of the total shear stress acting on the bed and banks is lost as “form drag” on large immobile grains and/or woody debris. As a result, it takes proportionally much more flow (and shear stress) to move the bed sediment than is actually available (see the discussion of “grain stress” in Wilcock and others 2008). The BAGS model does not fully correct for these effects, thus the model will tend to overpredict bed load transport rates in very high gradient streams. The alternative is to use a calibrated approach (Bakke and others 1999 or Wilcock 2001) taking as many bed load samples as possible at flows near the threshold for motion of surface-layer (framework) particles. As discussed in the next section, this flow typically occurs at approximately 60 to 70 percent of the bankfull discharge. If we follow this approach and selectively use observations from the highest flows to calibrate the transport equation, we get the relation shown by the dashed line in figure 10b. This relation matches the high-flow observations rather well, yet clearly under-predicts the bed load at lower flows. This may not be a cause for much concern; in typical gravel channels, flows less than about half of the bankfull discharge cumulatively carry a relatively small fraction of the total annual bed load (Emmett and Wolman 2001; Schmidt and Potyondy 2004; Torizzo and Pitlick 2004; Whiting and others 1999).

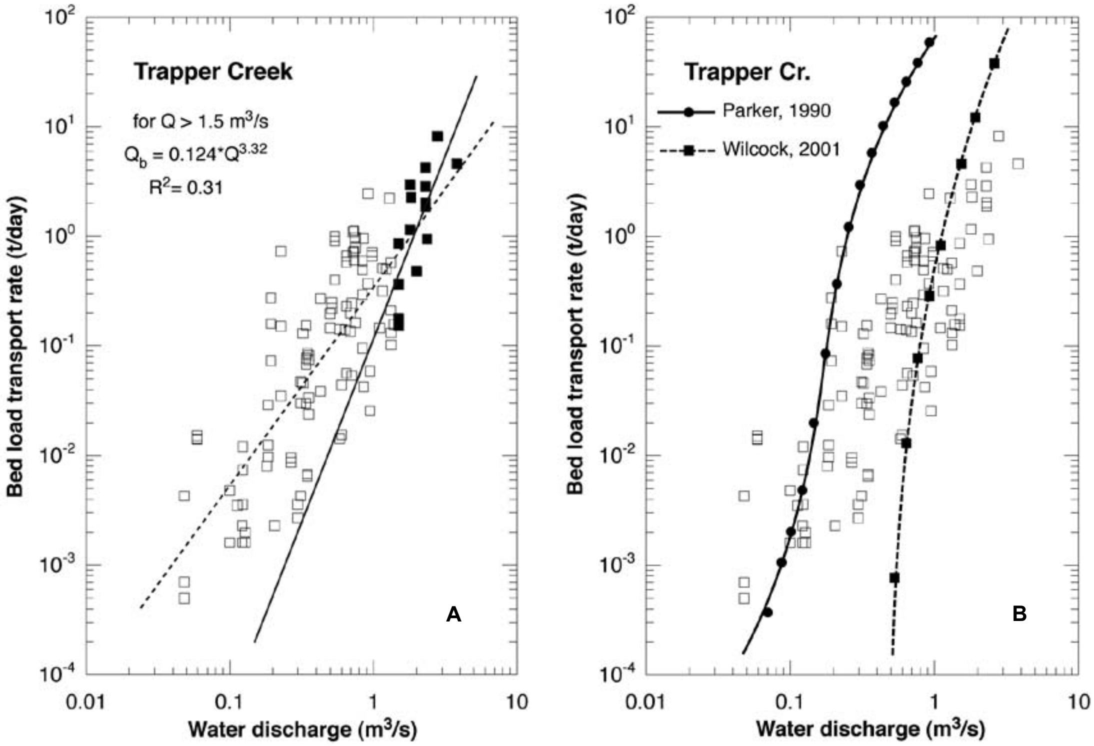


Figure 11. Observed (left) and modeled (right) relations between discharge and bed load transport rate, Trapper Creek, Idaho. In (A), the dashed line indicates the best-fit relation for the full data set, while the solid line is fit to a subset of the data, Q>1.5 m^3/s. In (B), the solid line indicates the surface-based equation of Parker (1990a) and the dotted line indicates the relation obtained with the calibrated approach of Wilcock (2001).

Assessing Model Output Without the Benefit of Bed Load Data

Significant time and effort were required to develop the data sets used in the preceding examples, and more often than not, the analyses required some tuning to achieve the best results. What are the chances then of developing accurate transport relations when you have no bed load data to serve as the basis for comparison? There isn't a definitive answer to this question; however, we can approach the problem by considering some of the conditions associated with transport measurements and observations elsewhere. We can start by asking two questions:

- Does the relation fail to predict any appreciable bed load transport at flows equal to the bankfull discharge?
- Does the relation predict appreciable, or even significant, bed load transport at flows less than about half of the bankfull discharge?

If the answer to either one of these questions is yes, the results of the transport calculation should be examined more carefully. We base this suggestion on the figure below, which summarizes results obtained by Mueller and others (2005) in the their analysis of bed load transport thresholds in natural channels. They compiled flow and bed load transport data for 45 gravel-bed streams and rivers in the western United States and Canada. For each of the 45 data sets, they plotted the relation between the dimensionless transport parameter, W^*, and dimensionless shear stress, τ^*. Then, following the procedure of Parker and others (1982), estimated the reference dimensionless shear stress, τ_r^*, corresponding to the reference transport rate of $W^* = 0.002$. Using the estimated values of τ_r and local hydraulic relations, they then computed the discharge associated with the reference transport rate and termed this the reference discharge, Q_r. Figure 12 shows a frequency distribution of Q_r, expressed as a ratio to the bankfull discharge, Q_{bf}. The distribution of Q_r is very nearly symmetric, with a median value of 0.67 (dashed line) and a mean of 0.68. This plot indicates that under natural (undisturbed) conditions, gravel-bed rivers typically begin mobilizing appreciable amounts of bed load when flows reach about two-thirds of the bankfull discharge. In only a few cases does the reference discharge lie above the bankfull discharge. Likewise, there are only a few cases where the reference discharge is less than ~30 percent of the bankfull discharge. Although conditions are likely to vary from one stream to another, we expect that the lower limit of appreciable transport will lie somewhere in this range.

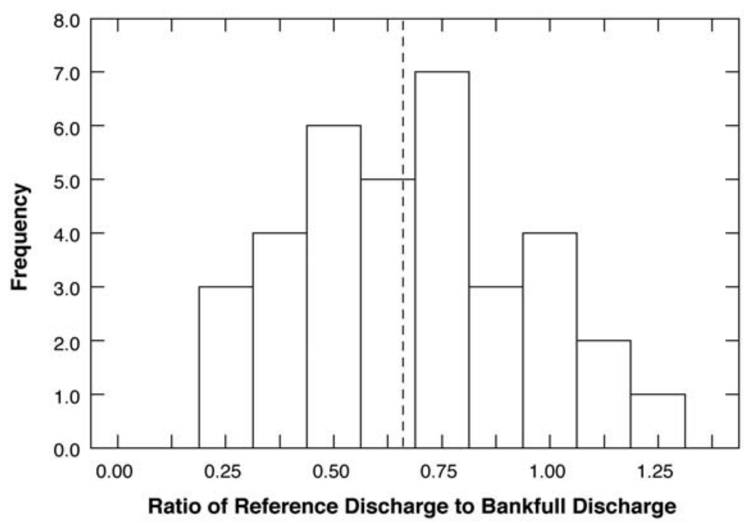


Figure 12. Frequency distribution of the ratio of the reference discharge for initiating bed load transport to the bankfull discharge, Q_r/Q_{bf}. Vertical dashed line indicates the median value (from Mueller and others 2005).

The discussion in the preceding paragraph leads to another question:

- What do we mean by appreciable transport? Is there a practical lower limit below which transport rates are so small that the loads can be considered negligible?

To answer this question, we should step back a bit and recall that the bed load transport relations used here compute finite loads for all flows, even relatively small ones. This is an intended feature of the models, rooted in the basic concept that entrainment and transport are probabilistic processes. Thus, even at very low flows, there is a small but finite probability that some bed load particles will move. This point is illustrated nicely by the bed load rating curves for Oak Creek and Halfmoon Creek presented earlier in figure 7. Using Halfmoon Creek as an example, the rating curve relation indicates that at a discharge of 2 m^3/s (~30 percent of the bankfull discharge), the creek carries a bed load of ~10^{-2} g/s integrated across the channel. How much sediment is that? We won't go through the details, but if we assume quartz-density sediment (ρ_s = 2.65 g/cm^3) and continuous transport, then a transport rate of 10^{-2} g/s equates to ~14 cm^3 of sediment per hour—a small hand full. If we maintain the same discharge for 8 days (the average annual frequency), the cumulative load is about 7 kg, which is much less than 1 percent of the total annual bed load (fig. 13). If we then double the discharge to 4 m^3/s (~60 percent of bankfull), the transport rate increases by two orders of magnitude to about 1.0 g/s. The bed load carried by that flow is still small—less than 2 percent of the total annual load (fig. 13)—but perhaps not so small that we would consider it to be negligible. The point here is clear. In many streams there may not be an absolute lower limit to bed load transport; however, it might be argued that there comes a point where the loads are so small that they can be considered negligible.

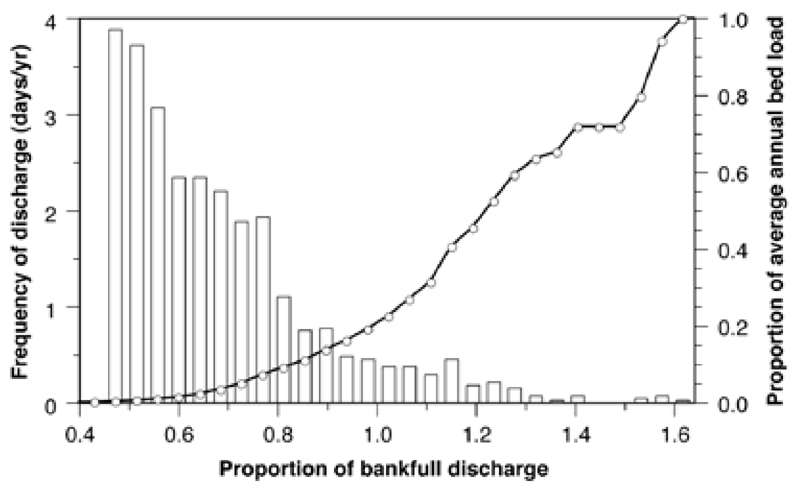


Figure 13. Relations between flow frequency and cumulative bed load transport for flows ranging from 0.4 to 1.6 times the bankfull discharge, Halfmoon Creek, CO.

The clearest guidance we can give is to suggest that at flows higher than about 2/3 of bankfull, bed load transport rates are likely to be non-negligible. The Parker transport models, for example, were developed under the assumption that the reference transport rate of W^* = 0.002 corresponds to discharges that are high enough to start mobilizing clasts within the armor layer. Mueller and others (2005) used the same assumption in their analysis, and listed the unit-width bed load transport rates, q_b, for each of the discharges corresponding to W^* = 0.002. Their results, summarized in figure 14, indicate that values of q_b at the reference discharge typically fall in the range between 0.001 and 0.005 kg/m/s. Based on this result, we suggest that at flows near reference discharge, predicted bed load transport rates should fall somewhere in the range from 0.001 to 0.005 kg/m/s.

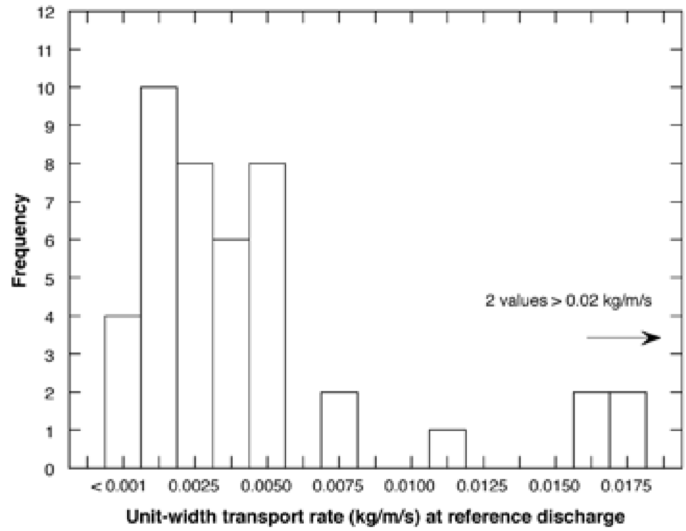


Figure 14. Frequency distribution of unit width bed load transport rates, q_b, at flows corresponding to the reference discharge, Q_r, in 45 gravel-bed streams and rivers in the United States and Canada (data from Mueller and others 2005).

Having posed the question of lower limits, we should consider the parallel question of upper limits. Specifically,

- Is there an upper limit to bed load transport or is there a way to constrain high-flow estimates of transport?

The short answer to the first part of this question is, yes, but in reality, any upper limit on bed load transport is not likely to be achieved in typical gravel bed streams, except during a debris flow. The second part of the question—constraining high-flow estimates—is a bit harder to answer. The clearest guidance we can give here is based on results from previous field studies. Table 2 provides a short list of transport rates measured at high flows in various river systems throughout the western United States. The sites are located in different regions, encompassing a range of conditions (hydrology, rock type, forest cover, and so forth). The maximum transport rates were selected from data given in the various reports or from graphs showing the individual transport relations. At the low end of the spectrum, there are several sites with maximum transport rates on the order of 10^{-2} kg/m/s. Several other sites have maximum bed load transport rates on the order of 10^{-1} kg/m/s. The highest value of 3.9 kg/m/s was measured on the North Fork of the Toutle River near Mount St. Helens, Washington. The headwaters of this river were severely disturbed during the May 1980 eruption of Mount St. Helens, thus transport rates measured downstream reflect unusual conditions within a highly erosive and unstable watershed. It is unlikely that gravel-bed streams in more stable settings will carry loads that high very often, if ever. Thus, if the predicted transport rates exceed 10 kg/m/s you should question the result. Reasonable values for maximum unit-width transport rates are more likely to fall in the range of 10^{-2} to 10^{-1} kg/m/s; however, this still leaves quite a bit of uncertainty and reinforces comments made earlier that a few bed load measurements will go a long way toward constraining such estimates.

Table 2. Examples of high bed load transport rates taken from measurements on gravel-bed streams and rivers in the western United States.

Site, reference	Drainage area (km^2)	Average channel gradient (m/m)	Bankfull discharge (m^3/s)	Unit stream —power at bankfull (watts/m^2)	Maximum measured bed load (kg/m/s)
Little Granite Cr[1]	55	0.0190	6.5	132.2	0.115
Main Fork Red River[2]	129	0.0059	9.3	44.4	0.017
Salmon R. nr. Obsidian[2]	243	0.0066	12.7	59.9	0.038
Boise River[2]	2154	0.0038	167	113.5	0.192
Selway River[2]	4955	0.0021	651	146.7	0.023
Snake River[3]	240766	0.0011	2607	153.8	0.173
Clearwater River[3]	24786	0.0006	2210	85.4	0.069
East Fork River[4]	466	0.0007	20	9.4	0.300
Jacoby Creek[5]	36	0.0062	32.6	165.2	0.400
Sagehen Creek[6]	27	0.0095	2.0	38.0	0.035
Virgin River[7]		0.0040	7.1	41.6	0.003
NF Toutle River[8]	736	0.0045	348		3.9

Data sources: (1) Ryan and Emmett (2002); (2) King and others (2004); (3) Jones and Seitz (1980); (4) Emmett (1980); (5) Lisle (1986); (6) Andrews (1994); (7) Andrews (2000); and (8) Pitlick (1992).

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