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Effects of sediment supply on surface textures of gravel-bed rivers

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Abstract. Using previously published data from flume studies, we test a new approach for quantifying the effects of sediment supply (i.e., bed material supply) on surface grain size of equilibrium gravel channels. Textural response to sediment supply is evaluated relative to a theoretical prediction of competent median grain size (D'_{50}). We find that surface median grain size (D_{50}) varies inversely with sediment supply rate and systematically approaches the competent value (D'_{50}) at low equilibrium transport rates. Furthermore, equilibrium transport rate is a power function of the difference between applied and critical shear stresses and is therefore a power function of the difference between competent and observed median grain sizes (D'_{50} and D_{50}). Consequently, we propose that the difference between predicted and observed median grain sizes can be used to determine sediment supply rate in equilibrium channels. Our analysis framework collapses data from different studies toward a single relationship between sediment supply rate and surface grain size. While the approach appears promising, we caution that it has been tested only on a limited set of laboratory data and a narrow range of channel conditions.

1. Introduction

Laboratory and field studies demonstrate that surface textures and gravel-bed rivers respond to changes in the rate of sediment supply (i.e., bed material supply) [*Leopold et al*., 1964; *Dietrich et al*., 1989; *Kinerson*, 1990; *Lisle and Hilton*, 1992, 1999; *Lisle et al*., 1993; *Rice*, 1994; *Nolan and Marron*, 1995]. When deprived of sediment, poorly sorted bed surfaces typically coarsen through size-selective winnowing of fine grains, as is common below dams [*Leopold et al*., 1964] and wood jams [*Rice*, 1994]. Winnowing results from transport rates in excess of sediment supply. Conversely, when inundated with sediment, bed load transport capacities may become overwhelmed, causing deposition of bed load material (typically fine-grained particles [*Leopold*, 1992]) and consequent reduction of surface grain size [*Rice*, 1994; *Nolan and Marron*, 1995].

On the basis of observed relationships between sediment supply and bed-surface texture, several methods have been proposed for qualitatively assessing sediment load (i.e., low, medium, and high) from inspection of surface grain size [*Dietrich et al*., 1989; *Lisle and Hilton*, 1992]. *Lisle and Hilton* [1999] recently demonstrated quantitative relationships between sediment yield and both the size and volume of fine sediment in pools. However, the observed relationships are sensitive to basin lithology and associated propensity for production of fine-grained sediment. In this paper we use data from previously published flume studies to test a new model for quantitatively determining rates of sediment supply from observed bed-surface grain size. We also use the model framework to further examine the effects of sediment supply on surface textures of gravel-bed rivers. Although surface textures can also respond to changes in the size distribution of the sediment supply [*Jackson and Beschta*, 1984; *Platts et al*., 1989; *Perkins*, 1989; *Knighton*, 1991; *Parker and Wilcock*, 1993; *Potyondy and Hardy*, 1994], we concentrate here on textural response to alterred sediment load of a fixed grain-size distribution. Furthermore, we focus on the special case of channels that are in equilibrium with their sediment supply.

2. Theory

Here we develop a theoretical relationship between sediment supply rate and bed-surface grain size of equilibrium channels. For equilibrium conditions the bed load transport rate is equal to the sediment supply rate ($q_b = q_s$). We define bed load transport rate as a power function of the difference between the applied and critical shear stresses

$$q_b = k(\tau' - \tau_c)^n = q_s, \tag{1}$$

where τ' is the bed shear stress (that portion of the total shear stress applied to the bed and responsible for sediment transport) [*Einstein and Barbarossa*, 1952; *Nelson and Smith*, 1989], τ_c is the critical shear stress for particle motion, and k and n are empirical values [*du Boys*, 1879; *O'Brien and Rindlaub*, 1934; *Meyer-Peter and Müller*, 1948; *Chien*, 1956; *Ashmore*, 1988; *Wathen et al*., 1995].

We define τ_c as the critical shear stress of the median bed-surface grain size (τ_{c50}). The median grain size (D_{50}) is related to its critical shear stress by the *Shields* [1936] equation, which is a force balance between the fluid stresses acting on a bed-surface particle and the resisting grain weight per unit area

$$\tau^*_{c50} = \tau_{c50}/(\rho_s - \rho)gD_{50}, \tag{2}$$

where ρ and ρ_s are the fluid and sediment densities, respectively, g is gravitational acceleration, and τ^*_{c50} is the dimensionless critical shear stress for incipient motion of D_{50}.

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Because the Shields equation is a force balance, it can be inverted to determine the competent median grain size for a given bed shear stress by defining that stress as the critical value

$$D'_{50} = \tau'/\tau^*_{c50}(\rho_s - \rho)g, \tag{3}$$

where D'_{50} is the competent median grain size for τ'. Consequently, (1) can be rewritten in terms of a difference of median grain sizes, rather than a difference of shear stresses

$$q_b = k_1(\tau' - \tau_c)^n = k_2(D'_{50} - D_{50})^n = q_s, \tag{4}$$

where D_{50} is the observed median grain size (proportional to τ_c) and τ^*_{c50} is assumed constant.

Several testable hypotheses for relating bed-surface grain size to sediment supply rate can be developed from (4). First, we hypothesize that if equilibrium transport rate is a power function of excess shear stress ($\tau' - \tau_c$), it also should be a function of the difference between D'_{50} and D_{50} (a surrogate for excess shear stress). If verifiable, rates of sediment supply could be determined from the difference of observed and competent median grain sizes. Second, we hypothesize that for given values of k, n, and τ', there should be an inverse relationship between sediment supply rate and τ_c (and thus D_{50}) for equilibrium transport.

3. Data Sources and Methods

Because it is difficult to demonstrate channel equilibrium in natural rivers, we tested the above hypotheses using data from flume studies where equilibrium transport could be verified. We used previously published data from laboratory studies that met the following criteria: (1) Experiments were conducted in sediment-feed flumes with surface textures, bed slope, and water surface slope allowed to equilibrate with an imposed discharge and sediment feed rate; (2) each investigator conducted experiments at several feed rates, using a constant size distribution; (3) experiments were allowed to run to a state of equilibrium transport (defined as equal rates of input and output); and (4) grain-size distributions of the equilibrium bed surface were reported. We found only four studies that met these criteria [*Parker et al*., 1982; *Kuhnle and Southard*, 1988, 1990; *Dietrich et al*., 1989; *Lisle et al*., 1993]. In two of the investigations [*Dietrich et al*., 1989; *Lisle et al*., 1993], equilibrium transport was additionally defined by similarity of input and output grain-size distributions. Furthermore, in these two studies the channel evolved in response to a progressive decrease in sediment supply rate. In contrast, the channel was reset for each run in the experiments by *Parker et al*. [1982] and *Kuhnle and Southard* [1988, 1990]. The compiled studies cover a broad range of channel slopes (0.0035–0.031), width/depth ratios (2–23), median grain sizes (2.3–12.9 mm), equilibrium transport rates (0.0028–1.07 kg m^{-1} s^{-1}), and bedform morphologies (plane bed, dune, and alternate bar) (Table 1). The primary sources of roughness in these studies are channel walls and form drag due to bed-surface grains and bed forms.

To test the relationships indicated by (4), values of the competent median grain size (D'_{50}) were calculated from (3) with τ^*_{c50} equal to 0.03, a conservative value for visually based studies of incipient motion [*Buffington and Montgomery*, 1997]. Moreover, τ^*_{c50} values near 0.03 are commonly reported from studies of bed load transport [e.g., *Parker and Klingeman*, 1982; *Andrews and Nankervis*, 1995], suggesting that 0.03 is an appropriate conservative value for use with bed load transport data in general. There is also historical precedent for choosing this value. In their well-known study of bed load transport, *Meyer-Peter and Müller* [1948, p. 54] propose $\tau^*_{c50} = 0.03$ as a lower limit for transport of mixed grain sizes.

Prediction of D'_{50} from (3) also requires estimation of the bed shear stress (τ', also known as the skin friction stress or fluid stress acting on the bed). The simplest and best known method for determining τ' is that developed by *Einstein* [1934, 1942, 1950; see also *Einstein and Barbarossa*, 1952]. However, the Einstein approach assumes a logarithmic velocity profile which is inappropriate for channels with grains that are large relative to the flow depth. Large grains cause significant particle form drag [*Andrews*, 1999] and development of a strongly nonlogarithmic velocity profile, producing bed stresses much smaller than would be predicted from a logarithmic velocity profile [*Wiberg and Smith*, 1991]. Relative roughness is a measure of potential particle form drag and is defined here as D_{84}/h, where D_{84} is the surface grain size for which 84% of the particles are smaller and h is flow depth. Values of D_{84}/h for the compiled laboratory studies range from 0.07 to 0.55, with an average of 0.28 (Table 1). Consequently, the Einstein method is not justified for the current data set. Instead, we use a more appropriate, albeit more complex, approach based on the work of *Wiberg and Smith* [1991].

Wiberg and Smith [1991] developed a model for simultaneous solution of the vertical velocity profile and the form drag due to bed-surface grains; simultaneous solution is required because the two influence one another. Unlike the Einstein approach, the form of the velocity profile is not specified a priori, allowing more accurate determination of fluid forces in conjunction with site-specific measures of bed material properties. We modified the *Wiberg and Smith* [1991] model to include other sources of roughness observed in the compiled studies. We calculated the bed shear stress (τ') in (3) as a reach-average depth-slope product corrected for wall effects [*Houjou et al*., 1990], particle form drag [*Wiberg and Smith*, 1991], and bed form drag where present [*Nelson and Smith*, 1989]

$$\tau_f(z) = \tau_T(z) - [\tau_w(z) + \tau_p(z) + \tau_{bf}(z)], \tag{5}$$

where τ_f is the fluid stress (equal to τ' at the bed), τ_T is the total stress, τ_w is the stress associated with wall roughness, τ_p is the stress caused by particle form drag, τ_{bf} is the stress associated with bed form drag, and z is height above the bed. A finite difference model is used to simultaneously determine each of these reach-average shear stress components in conjunction with the reach-average vertical velocity profile [*Houjou et al*., 1990; *Wiberg and Smith*, 1991].

Following the approach of *Wiberg and Smith* [1991], the vertical gradient of the downstream velocity ($\partial u/\partial z$) is expressed in terms of the above stresses as

$$\frac{\partial u}{\partial z} = \frac{1}{L}\left(\frac{\tau_f}{\rho}\right)^{0.5} = \frac{1}{L}\left[\frac{\tau_T - (\tau_w + \tau_p + \tau_{bf})}{\rho}\right]^{0.5}, \tag{6}$$

where L is the length scale of turbulent mixing (as defined by *Wiberg and Smith* [1991]). Rearranging terms and integrating yields

$$u(z) = \int_{z_0}^{z} \frac{1}{L}\left[\frac{\tau_T - (\tau_w + \tau_p + \tau_{bf})}{\rho}\right]^{0.5} dz, \tag{7}$$

Table 1. Equilibrium Channel Characteristics of the Compiled Studies

Source	Run	Bed Morphology	h, m	w, m	w/h	S*	D_{84}/h	τ_0, Pa	τ'/τ_0	τ_p/τ_0	τ_{bf}/τ_0†	τ_w/τ_0	ρ_s, kg m^{-3}‡	$q_b = q_s$, kg m^{-1} s^{-1}	D_{50f}, mm	σ_f (ϕ)	D_{50}, mm§,‖	σ (ϕ)	D'_{50}, mm
Parker et al. [1982]	2	plane bed	0.0284	0.457	16	0.0118	0.30	3.29	0.75	0.22	NA	0.03	2600	0.0036	2.0	1.20	4.5	1.05	5.21
	4	weak alternate bar	0.0353	0.457	13	0.0102	0.24	3.53	0.75	0.20	?	0.04	2600	0.0091	2.0	1.20	4.7	1.16	5.66
Kuhnle and Southard [1988, 1990]	L3	low-amplitude dunes? (bed load sheets?)¶	0.046	0.15	3	0.024	0.50	10.83	0.70	0.19	NA	0.11	2600	0.0340	3.03	1.40	11.5	1.19	16.02
	H1	low-amplitude dunes? (bed load sheets?)¶	0.046	0.53	12	0.019	0.54	8.57	0.75	0.21	NA	0.04	2600	0.0340	2.95	1.48	12.9	1.50	13.75
	H2	low-amplitude dunes? (bed load sheets?)¶	0.045	0.15	3	0.024	0.55	10.59	0.70	0.19	NA	0.11	2600	0.0340	2.95	1.48	12.9	1.85	15.70
	H3	low-amplitude dunes (bed load sheets)¶	0.074	0.15	2	0.015	0.23	10.89	0.65	0.18	NA	0.17	2600	0.0980	2.95	1.48	8.9	1.24	15.05
	H5	dunes	0.069	0.15	2	0.021	0.11	14.21	0.65	0.17	0.005	0.18	2600	1.073	2.95	1.48	2.6	1.31	19.61
Dietrich et al. [1989]**	17.4	low-amplitude dunes (bed load sheets)¶	0.102	0.300	3	0.0052	0.07	5.20	0.67	0.17	NA	0.16	2650	0.0290	3.63	1.29	3.74	1.42	7.17
	6.1	plane bed	0.103	0.300	3	0.0046	0.07	4.65	0.68	0.16	NA	0.16	2650	0.0102	3.63	1.29	4.26	0.97	6.47
	1.7	plane bed	0.113	0.300	3	0.0035	0.07	3.88	0.67	0.14	NA	0.18	2650	0.0028	3.63	1.29	4.85	1.24	5.37
Lisle et al. [1991, 1993]	11.7	alternate bar	0.011	0.241	22	0.031	0.37	3.35	0.62	0.20	0.15	0.02	2650	0.0337	1.4	1.02	2.3	0.84	4.31
	12.3	alternate bar	0.010	0.229	23	0.027	0.44	2.65	0.68	0.23	0.06	0.03	2650	0.0114	1.4	1.02	2.8	0.66	3.71
	13	alternate bar	0.011	0.237	22	0.028	0.49	3.02	0.66	0.22	0.10	0.03	2650	0.0034	1.4	1.02	3.7	0.59	4.11

The following are reach-average values: h (flow depth), w (flow width), S (slope), D_{84} (surface grain size for which 84% of the particles are smaller), τ_0 (total boundary shear stress), τ' (fluid shear stress at the bed), τ_p (stress due to particle form drag), τ_{bf} (stress due to bed form drag), τ_w (stress associated with channel walls), D_{50} (median bed-surface grain size), σ (graphic standard deviation [*Folk*, 1974] of surface distribution), and D'_{50} (competent median surface grain size, equation (3)). Other values are defined as follows: q_b, bed load transport rate; q_s, sediment supply rate; D_{50}, median grain size of the sediment feed; σ_f, graphic standard deviation of feed distribution; and ρ_s, sediment density. NA is not applicable.

* All slopes are water-surface values, except those of *Kuhnle and Southard* [1988, 1990], which are bed-surface values.

† We assumed unseparated flow for the high-amplitude bars in *Lisle et al.*'s [1993] experiments (bed form drag coefficient $C_d = 0.84$) and used values of A_x/A_b reported in their Table 1. We assumed separated flow for the low-amplitude dunes in *Kuhnle and Southard*'s [1988] run H5 ($C_d = 0.21$), with dune amplitude equal to 1 cm and wavelength equal to 60 cm. We did not calculate form drag caused by bed load sheets because amplitude and wavelength data were incomplete or unavailable. However, these features are unlikely to offer much resistance because of small amplitudes (2–4 mm for those of *Kuhnle and Southard* [1988] and 1–2 grain diameters for those of *Dietrich et al.* [1989]) and wavelengths typically many times greater than amplitude [*Whiting et al.*, 1988]. No data were available to calculate form drag caused by the weak alternate bars in *Parker et al.*'s [1982] run 4.

‡ *Kuhnle and Southard*'s [1988, 1990] H runs are mixtures of light and heavy minerals, predominantly composed of quartz (97% by weight, $\rho_s = 2600$ kg m^{-3}, $D = -5$ to 3.5 ϕ), with monir amounts of magnetite (0.9% by weight, $\rho_s = 5200$ kg m^{-3}, $D = 0.5$ to 3 ϕ), lead (1.2% by weight, $\rho_s = 11400$ kg m^{-3}, $D = 0.5$ to 1.5 ϕ), and tungsten (0.9% by weight, $\rho_s = 19300$ kg m^{-3}, $D = 0.5$ to 3 ϕ).

§ *Parker et al.* [1982] and *Kuhnle and Southard* [1988, 1990] sampled surface sizes with a clay piston, producing semivolumetric samples. *Dietrich et al.* [1989] and *Lisle et al.* [1993] sampled surface sizes using gridded point counts.

‖ D_{50} values for *Kuhnle and Southard* [1988] were estimated from their Figure 14, and those of *Lisle et al.* [1993] were estimated from their Figure 1.

¶ Morphology is that of *Iseya and Ikeda* [1987] and *Whiting et al.* [1988].

** Grain-size data are from *Kirchner et al.* [1990].

where z_0 is the roughness length scale of the bed and $u = 0$ at $z = z_0$. Equation (7) is integrated numerically, with the stress components defined in terms of u and z, allowing iterative, simultaneous solution of both the stresses and the velocity profile [*Wiberg and Smith*, 1991].

The reach-average total shear stress is written as a depth-slope product that varies as a linear function of height above the bed

$$\tau_T(z) = \tau_0 \left(1 - \frac{z}{h}\right) = \rho g h S \left(1 - \frac{z}{h}\right), \tag{8}$$

where τ_0 is the total boundary shear stress ($\rho g h S$) and h and S are the reach-average flow depth and water surface slope, respectively.

Form drag caused by bed-surface grains, bed forms, or any other object in the flow can be expressed as

$$F_d = \frac{\rho}{2} C_d \langle u^2 \rangle A_x, \tag{9}$$

where F_d is the drag force that results from pressure differences across the object, C_d is the drag coefficient, A_x is the flow-perpendicular area of the object, and $\langle u^2 \rangle$ is the square of the reference velocity averaged over A_x. The reference velocity is defined as the velocity that would exist in the absence of the object of interest. Dividing the drag force by the flow-parallel area of the object (A_b) yields the associated drag stress

$$\tau_d = \frac{\rho}{2} C_d \langle u^2 \rangle \frac{A_x}{A_b}. \tag{10}$$

The stress due to grain form drag is defined according to *Wiberg and Smith* [1991] as the particle drag force per unit area summed over all grain sizes (smallest to largest, D_i to D_M) at each level of z

$$\tau_p(z) = \sum_{m=i}^{M} \left(\frac{F_d}{A_b}\right)_m \tag{11a}$$

$$= \frac{\rho}{2} \sum_{m=i}^{M} \left(C_d \langle u^2 \rangle \frac{A_x}{A_b} c \right)_m \tag{11b}$$

$$= \frac{\rho}{2} \sum_{m=i}^{M} \left[\left(C_d c \frac{A_x}{A_b} \right)_m \int_{z_0}^{z_m} u^2 \, dz/(z_m - z_0) \right], \tag{11c}$$

where c_m is the concentration of a component grain size and z_m is the vertical height of that particle above the bed. We assume elliptical grains with a *Corey* [1949] shape factor of 0.6 (a reasonable value both for natural sediments and those used in the laboratory studies investigated here). The shape factor is defined as $D_S/\sqrt{D_I D_L}$, where the subscripts S, I, and L indicate short, intermediate, and long axes, respectively. We assume that the grains are oriented with their short axes vertical and that the intermediate and long diameters are equal to one another ($D_S = 0.6 D_I = 0.6 D_L$). Using the nominal grain diameter ($(D_S D_I D_L)^{1/3}$), C_{d_m} is approximated by the drag coefficient for a sphere, which is a function of particle Reynolds number [e.g., *Rouse*, 1946]. While developed for isolated spheres well away from channel boundaries, this relationship between drag coefficient and Reynolds number approximates that of spheres resting on a bed of like grains [*Coleman*, 1967].

Because the size distribution of the competent bed surface is unknown, we approximated c_m using a lognormal grain-size distribution

$$c_m = \frac{c_b}{\sigma \sqrt{2\pi}} \exp\left[-\frac{1}{2} \left(\frac{\phi_m - \phi_{50}}{\sigma} \right)^2 \right], \tag{12}$$

where c_b is the maximum concentration of bed-surface particles (set equal to 0.6), ϕ_m and ϕ_{50} are the component and median grain sizes, respectively, expressed in ϕ units ($\phi = -\log_2 D$, where D is in millimeters [*Krumbein*, 1936]) and σ is the standard deviation of particle sizes in ϕ. We chose the smallest reported value of σ in each investigation as a conservative estimate for the competent bed surface, with the range of grain sizes limited by those used in each study. Here ϕ_{50} is determined through simultaneous solution of (3) and (5) (discussed further below).

Similarly, the stress associated with drag due to two-dimensional bed forms is defined by *Nelson and Smith* [1989] as

$$\tau_{bf}(z) = \frac{\rho}{2} C_d \langle u^2 \rangle \frac{A_x}{A_b} = \frac{\rho}{2} C_d \frac{A_x}{A_b} \int_{z_0}^{z_{bf}} u^2 \, dz/(z_{bf} - z_0), \tag{13}$$

where z_{bf} is the height of the bed form. Here the drag coefficient is 0.21 if flow separates around the bed form and is 0.84 if flow remains unseparated [*Smith and McLean*, 1977].

Wall effects include both momentum diffusion caused by proximity of walls (i.e., width-to-depth effects) [*Leighly*, 1932; *Parker*, 1978] and differences between bed and wall roughness [*Einstein*, 1934; 1942; *Houjou et al.*, 1990]. Using data from *Houjou et al.*'s [1990] study of planar rectangular channels, the ratio of wall to bed stress (τ_w/τ') can be expressed as a power function of the width/depth ratio (w/h) and the ratio of bed to wall roughness (z_{0_b}/z_{0_w})

$$\frac{\tau_w}{\tau'} = 2.55 \left(\frac{w}{h}\right)^{-1.1} \left(\frac{z_{0_b}}{z_{0_w}}\right)^{-0.21} \qquad R^2 = 0.99, \tag{14}$$

where all values are cross-sectional and reach averages. Rearranging (14) yields an expression for τ_w as a function of τ', w/h, and z_{0_b}/z_{0_w}

$$\tau_w = 2.55 \tau' \left(\frac{w}{h}\right)^{-1.1} \left(\frac{z_{0_b}}{z_{0_w}}\right)^{-0.21}. \tag{15}$$

The effective roughness of the bed in planar channels is $z_{0_b} \approx 0.1 D_{84}$ [*Whiting and Dietrich*, 1990; *Wiberg and Smith*, 1991]. This definition of z_{0_b} includes both grain skin friction and particle form drag [*Wiberg and Smith*, 1991]. The effective bed roughness in channels with bed forms is found by matching the inner and outer velocity profiles at the height of the bed form (z_{bf}) [*Nelson and Smith*, 1989], yielding

$$z_{0_b} = z_{bf} \left(\frac{z_{bf}}{0.1 D_{84}} \right)^{-\sqrt{(\tau' + \tau_p)/(\tau' + \tau_p + \tau_{bf})}}. \tag{16}$$

We define the wall roughness from *Nikuradse*'s [1950] equation for hydraulically smooth flow, appropriate for the smooth-walled flumes studied here,

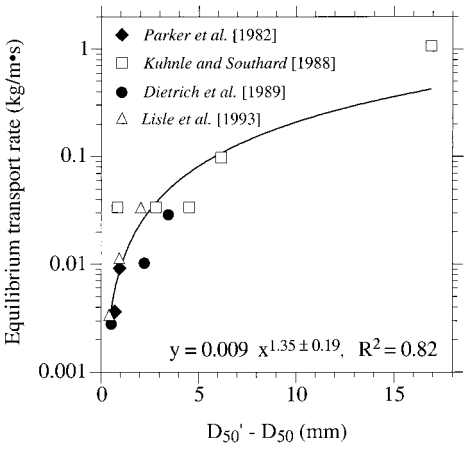

Figure 1. Equilibrium transport rate ($q_b = q_s$) as a function of the difference between competent median grain size (D'_{50}) and the observed value (D_{50}).

$$z_{0_w} = \nu/9u^*_w, \tag{17}$$

where ν is the kinematic velocity and u^*_w is the shear velocity at the wall ($\sqrt{\tau_w/\rho}$).

The competent median grain size (D'_{50}) is determined from simultaneous, iterative solution of (3), (5), (8), (11), (13), and (15). Because D'_{50} is a function of bed shear stress, which is, in turn, a function of the surface grain-size distribution and consequent particle form drag, the competent median grain size and the bed shear stress must be solved together. We emphasize that the competent median grain size is a hypothetical value, the occurrence of which depends on the local supply and caliber of sediment. Channel characteristics and calculated values for the compiled studies are presented in Table 1.

4. Use of Surface Grain Size to Infer Sediment Supply

Data from the compiled laboratory studies support the form of (1) quite well, demonstrating that equilibrium transport ($q_b = q_s$) is a power function of the difference between the applied and critical shear stresses. The equilibrium transport equation for these data is $q_b = 0.025(\tau' - \tau_c)^{1.35 \pm 0.19}$ with $R^2 = 0.82$. Although the exponent of this equation is within the range of typically reported values (1.3–1.8 [*O'Brien and Rindlaub*, 1934; *Meyer-Peter and Müller*, 1948; *Chien*, 1956; *Ashmore*, 1988]), both the coefficient and exponent of the relationship are sensitive to investigator choice of (1) the method for determining τ_c, (2) the characteristic grain size represented by τ_c (i.e., the particular grain-size percentile and the population it is drawn from (surface, subsurface, or load)), and (3) the method of correcting τ' for channel roughness. Depending on the above choices, values of the coefficient k vary by an order of magnitude (0.003–0.06), while values of the exponent n vary from 1.2 to 2.1 for this data set [*Buffington*, 1998].

Because equilibrium transport is a power function of excess shear stress, it follows that it is also a power function of the competent and observed median grain sizes (surrogates for the applied and critical shear stresses, (4)), as demonstrated by Figure 1. Plotted in this fashion, the data collapse toward a single function despite differing (1) bed shear stress, (2) reach-average median surface grain size, (3) caliber of supplied sediment, and (4) sorting of supplied sediment in the four studies (Table 1). Previous methods for assessing textural response to sediment supply resulted in a family of curves that were dependent on study-specific ratios of bed shear stress to critical shear stress of the bed load material [*Dietrich et al*., 1989; *Lisle et al*., 1993]. The general collapse of the data toward a single function suggests that the analysis framework used here provides a unifying relationship between sediment supply rate and surface grain size in equilibrium channels.

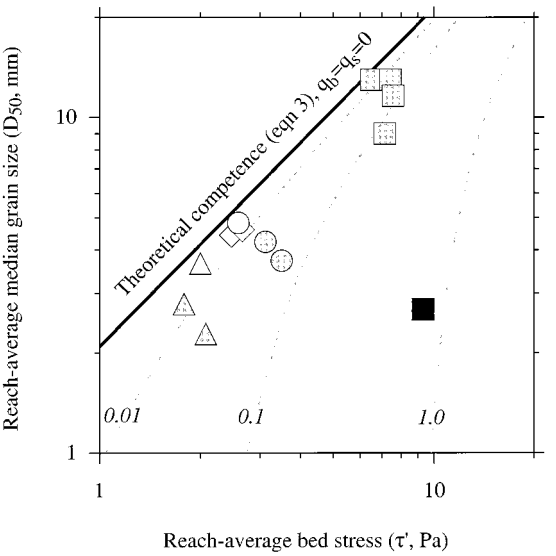

Figure 2. Observed median surface grain size (D_{50}) versus bed shear stress (τ') for experiments by *Parker et al*. [1982] (diamonds), *Kuhnle and Southard* [1988] (squares), *Dietrich et al*. [1989] (circles), and *Lisle et al*. [1993] (triangles). The data are stratified by equilibrium transport rates ($q_b = q_s$) of <0.01, <0.1, and >1 kg m^{-1} s^{-1} shown as open, shaded, and solid, respectively; specific equilibrium transport rates for each data point are listed in Table 1. The dashed lines are predictions of D_{50} for equilibrium transport rates of 0.01, 0.1, and 1 kg m^{-1} s^{-1} based on the curve fit of Figure 1, assuming $\tau^*_{c50} = 0.03$ and $\rho_s = 2650$ kg m^{-3}. The bold solid line is the theoretical prediction of competent median grain size (D'_{50}, see equation (3)).

5. Effects of Sediment Supply on Surface Grain Size

Figure 2 compares the observed median grain size to the predicted competent value as a function of bed shear stress and sediment supply rate. The data are stratified by equilibrium transport rates of <0.01, <0.1, and >1 kg m^{-1} s^{-1}, with specific values listed in Table 1; contours of predicted grain size for these transport rates also are shown based on the curve fit of Figure 1. The data of *Lisle et al*. [1993] and *Kuhnle and Southard* [1988] show a wide range of median surface grain sizes for a relativley small range of bed shear stresses, with D_{50} varying inversely with sediment supply as hypothesized from (4). Furthermore, the compiled data show that median surface grain size systematically approaches the competent value at low equilibrium transport rates, as predicted from (4). This equation indicates that the observed surface grain size should approach the theoretical competent value as equilibrium transport rates go to zero.

6. Discussion and Conclusions

Our analysis demonstrates that surface textures respond to changes in sediment supply rate, as documented by *Dietrich et al*. [1989] and *Lisle et al*. [1993]. Imbalances between rates of sediment supply and bed load transport cause size-selective deposition or erosion that alters bed roughness which, in turn, likely alters both the critical shear stress for grains traveling over the bed [*Kirchner et al*., 1990; *Buffington et al*., 1992; *Johnston et al*., 1998] and the effective bed shear stress [*Naot*, 1984; *Wiberg and Smith*, 1991]. These responses feed back on one another, partially counteracting local imbalances between sediment supply and transport capacity [*Dietrich et al*., 1989; *Lisle et al*., 1993]. For example, transport capacities overwhelmed by high sediment supplies should cause fine-sediment deposition, creating smoother bed surfaces with lower intergranular friction angles and higher bed stresses, both of which will promote increased transport rates across that surface (equation (1)), reducing the difference between supply and transport rates.

Parker and Klingeman [1982, p. 1422] hypothesized that surface texture and degree of armoring in gravel-bed channels are controlled by the ratio of bed stress to critical stress (an alternative expression for excess shear stress). They suggested that "when stresses are well above critical stress, mobility differences on a grain-by-grain basis are reduced; the differences disappear asymptotically. Thus a coarse pavement should not form. The low-stress range for which pavement is required is typical of gravel bed streams; the high-stress range where it should not occur is typical of sand bed streams." Subsequent laboratory and field studies corroborate their hypothesis [*Parker et al*., 1982; *Wilcock and Southard*, 1989; *Pitlick*, 1992]. We further suggest that bed-surface texture represents a feedback between rates of both sediment supply and bed load transport (the latter being a function of surface roughness, excess shear stress, and availability of transportable sediment).

Although bed-surface texture may respond to rates of both sediment supply and bed load transport (i.e., excess shear stress), our use of equilibrium conditions renders these two factors equal and inseparable, allowing interpretation of equilibrium rates of sediment supply from inspection of bed-surface grain size. The analysis presented here extends previous studies of the relationship between sediment supply rate and surface grain size [*Dietrich et al*., 1989; *Lisle et al*., 1993] and shows that there may be a single relationship when sediment supply is related to the difference between D'_{50} and D_{50}. We find that this difference is an indirect measure of excess shear stress ($\tau' - \tau_c$) and is well correlated with rates of equilibrium transport across different studies (Figure 1). Although there are investigation-specific trends within the data, our analysis framework produces a reasonable collapse of the compiled data.

However, our analysis assumes a constant value of τ^*_{c50} and so does not fully describe textural response to altered rates of equilibrium transport. Dimensionless critical shear stress (τ^*_c) will vary with site-specific differences in bed material properties (e.g., grain shape, rounding, protrusion, imbrication, and distribution of intergranular friction angles). Moreover, τ^*_c values will covary with changes in bed-surface grain size and sorting that occur in response to altered rates of bed load transport or sediment supply. For example, *Church et al*. [1998] observed temporal changes in effective values of τ^*_{c50} as bed surfaces coarsened and as bed material structure changed in response to elimination of sediment supply. Temporal changes in bed-surface grain size and sorting will also affect friction angle distributions and consequent values of τ^*_c [*Kirchner et al*., 1990; *Buffington et al*., 1992; *Johnston et al*., 1998]. Our constant value of $\tau^*_{c50} = 0.03$ is generally appropriate for nonimbricated plane beds, but higher values of τ^*_c are advised for channels with more copmlex bed material structures (e.g., imbricated grains, particle clusters, or stone cells) that reduce particle mobility [*Church et al*., 1998].

Despite the use of a constant value of τ^*_{c50} we find a strong relationship between surface grain size and equilibrium transport rate for the data examined here. On the basis of our findings we suggest that rates of sediment supply in equilibrium channels can be determined from the difference between observed median grain size and the predicted competent value (Figure 1). Regardless of whether the caliber of the observed surface grain size reflects textural response to excess shear stress, sediment loading (Figure 2), or the particular size distribution of the supply, the difference between predicted and observed grain size is a measure of excess shear stress and is therefore a measure of both q_b and q_s for equilibrium channels.

Assessing rates of sediment supply in natural gravel-bed rivers using our approach requires conditions of quasi-equilibrium. For equilibrium channels with well-developed floodplains and self-formed beds (i.e., all sizes mobile at typical high flows), D'_{50} should be predicted from the bank-full bed stress. An expedient approach for applying Figure 1 is to confine its use to hydraulically simple, plane-bed reaches where correction for bank effects and particle form drag may be the only roughness correction needed. Hydraulically simple reaches are recommended because the grain-size response to sediment supply observed in our analysis is fairly small (~17 mm over 3 orders of magnitude of sediment loading) and may be overwhelmed by parameter uncertainty in formulas for stress correction of hydraulically complex reaches. *Lisle and Hilton* [1999] also report fairly small magnitudes of textural response to sediment loading in gravel-bed rivers of northern California and southern Oregon. They found that the D_{50} of fine pool material decreases approximately 9 mm over 3 orders of magnitude of sediment yield. The sensitivity of equilibrium transport rates to small changes in grain size also highlights the importance of accurate grain-size sampling, insuring a high degree of confidence for the grain-size statistics (further discussion of sediment sampling procedures and accuracy is given by *Church et al*. [1987], *Rice and Church* [1996], and *Buffington and Montgomery* [1999]).

While Figure 1 may prove useful for assessing rates of local bed load sediment supply, it is cautioned that our results are based on a limited set of flume data with small grain sizes ($D_{50} < 13$ mm) and channels not entirely self-formed; except for the study of *Lisle et al*. [1993] the laboratory channels used in our compilation were constrained to specific widths. Furthermore, the caliber of sediment supply was held constant in each experiment, and although the data collapse fairly well into a single function despite differing size distributions of sediment supply, the limited amount of available data does not allow a full exploration of the effects of these parameters. For example, everything else being equal, bed surface response to a supply of fine-grained sediment may be very different than to a supply of coarser-grained material. Use of a constant value of τ^*_{c50} also may cause errors in calculated values of D'_{50} and corresponding estimates of equilibrium transport rate (as dis-

cussed above). Because of these concerns, further investigation of our findings is warranted before they are applied as a land management tool. Moreover, the condition of near-equilibrium transport may not be a realistic assumption for actively disturbed landscapes, such as watersheds that have been intensely clear-cut and are experiencing high sediment loads due accelerated landslide frequency, road runoff, or destabilized river banks. Where it is known that channels are aggrading, our proposed method provides a minimum estimate of local sediment supply rate; the difference between predicted and observed D_{50} can provide an estimate of the bank-full q_b, which must be less than q_s for an aggrading system.

The requirement of equilibrium transport for use of our approach motivates an interesting question about natural channels: What is the time for bed-surface adjustment to perturbations of sediment supply or transport capacity? That is, how long does it take for surface grain size to equilibrate with imposed channel hydraulics and sediment load? Gravel-bed channels occur in a wide range of environments that are characterized by different magnitudes and frequencies of discharge events and sediment inputs, offering a range of timescales for channel adjustment to these perturbations [*Pitlick*, 1994]. In arid regions, channels experience infrequent, intense floods of short duration (on the scale of several hours) that may not offer sufficient time for morphologic adjustment to the imposed discharges and sediment loads [*Laronne and Reid*, 1993]. In mountain drainage basins of the west coast of North America, hydraulic discharge is typically perennial, with rainfall-driven flood events occurring one or more times a year and having durations of the order of one or more days, allowing significantly more time for morphologic adjustments. In intermontaine regions of the United States, flood flows typically result from spring snowmelt and are characterized by long, gradually varying discharges that occur of the order of tens of days to months and allow even more time for morphologic adjustments and attainment of quasi-equilibrium. To put these environmental differences in context, the 100 year flood may be less than 2 times the mean annual flow in snowmelt basins, while it can be more than 10 times the mean annual flow in watersheds with flashy, storm-driven hydrographs [*Pitlick*, 1994].

The likelihood of channel morphology equilibrating with the imposed hydrology and sediment load depends on the shape and duration of the local hydrograph, as well as the timescale of adjustment for the morphologic feature of interest. Moreover, alluvial channels are nonlinear dynamic systems that can exhibit a variety of responses to perturbations of hydraulic discharge or sediment supply. For example, many alluvial channels are free to adjust their width, depth, slope, bed form morphology, channel pattern, and bed-surface texture in response to changes in discharge and sediment supply (see review by *Montgomery and Buffington* [1998]). Several morphologic adjustments may occur simultaneously for a given perturbation, with many responses feeding back on one another. The specific morphologic response to a particular channel perturbation depends on local channel characteristics, geomorphic setting, and historical events within the watershed. Nevertheless, changes in bed-surface textures may be an active and potentially rapid, first-order response to altered sediment supply or hydraulic discharge. Consequently, observing changes in bed-surface texture may be a more sensitive and effective means of monitoring channel response to altered sediment supply and hydraulic discharge over short timescales. Over longer timescales, other morphologic features may be better indicators of channel response, particularly where textural changes are short-lived or counteracted by subsequent, larger-scale morphologic adjustments.

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