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Reviews of Geophysics

Key Points:
- Transport capacity has emerged as a concept several times in geomorphology
- Flow transport-surface change feedbacks limit flow power-capacity relationships
- Capacity relationships do not scale in space or time as transport is stochastic

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The concept of transport capacity in geomorphology

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Abstract The notion of sediment-transport capacity has been engrained in geomorphological and related literature for over 50 years, although its earliest roots date back explicitly to Gilbert in fluvial geomorphology in the 1870s and implicitly to eighteenth to nineteenth century developments in engineering. Despite cross fertilization between different process domains, there seem to have been independent inventions of the idea in aeolian geomorphology by Bagnold in the 1930s and in hillslope studies by Ellison in the 1940s. Here we review the invention and development of the idea of transport capacity in the fluvial, aeolian, coastal, hillslope, débris flow, and glacial process domains. As these various developments have occurred, different definitions have been used, which makes it both a difficult concept to test, and one that may lead to poor communications between those working in different domains of geomorphology. We argue that the original relation between the power of a flow and its ability to transport sediment can be challenged for three reasons. First, as sediment becomes entrained in a flow, the nature of the flow changes and so it is unreasonable to link the capacity of the water or wind *only* to the ability of the fluid to move sediment. Secondly, environmental sediment transport is complicated, and the range of processes involved in most movements means that simple relationships are unlikely to hold, not least because the movement of sediment often changes the substrate, which in turn affects the flow conditions. Thirdly, the inherently stochastic nature of sediment transport means that any capacity relationships do not scale either in time or in space. Consequently, new theories of sediment transport are needed to improve understanding and prediction and to guide measurement and management of all geomorphic systems.

1. Introduction

The notion of sediment-transport capacity has been engrained in geomorphological and related geophysical literature for over 50 years. However, the definition has not always been consistent across the discipline. One reason for this inconsistency has been the tendency in recent decades for increasing specialization in studies relating to particular process domains. Together with the proliferation of literature, specialization means that only rarely are evaluations made of linkages between concepts in different process domains. This trend toward increasingly focussed specialization, however, contrasts with developments in complexity theory, Earth Systems Science, and integrated management of the environment that all emphasize the benefits of holistic approaches. In this review, we will first examine how the notion of transport capacity has been developed and applied in the various branches of geomorphology in order to evaluate commonalities and divergences in approach, in relation to conceptual developments and empirical support. In doing so, we assess the effectiveness of the concept by using advances across the different process domains to inform and interpret each other and thereby provide an overall critique of the concept. We take a critical realist approach [*Richards*, 1990; *Sayer*, 2000] to assess whether the use of the concept, both within a single process domain and between different process domains, enables the explanation of observations both in the field and laboratory. In other words, if the definitions of transport capacity that have been developed are real—mechanistic—characterizations of how sediment is transported in geomorphic systems, they should be able to explain observed phenomena. If not, there needs to be a reevaluation of whether it is the concept (or concepts) that are problematic or the experimental approaches and data that are flawed.

GK Gilbert was central to the development of the concept of transport capacity in geomorphology. In his *Report on the Geology of the Henry Mountains* [*Gilbert*, 1877] he noted "Transportation is favored by increasing water supply as greatly as by increasing declivity. When the volume of a storm increases, it becomes at the same time more rapid, and its transporting capacity gains by the increment to velocity as by the increment to volume. Hence the increase in power of transportation is more than proportional to the increase in volume" [*Gilbert*, 1877, p. 98].

Based on a steady state conceptualization of river form, and an early formulation of a concept of stream power, he goes on to suggest: "Let us suppose that a stream endowed with a constant volume of water, is at some point continuously supplied with as great a load as it is capable of carrying. For so great a distance as its velocity remains the same, it will neither corrade (downward) nor deposit, but will leave the grade of its bed unchanged. But if in its progress it reaches a place where a less declivity of bed gives a diminished velocity, its capacity for transportation will become less than the load and part of the load will be deposited. Or if in its progress it reaches a place where a greater declivity of bed gives an increased velocity, the capacity for transportation will become greater than the load and there will be corrasion of the bed. In this way a stream which has a supply of débris equal to its capacity, tends to build up the gentler slopes of its bed and cut away the steeper. It tends to establish a single, uniform grade" [*Gilbert*, 1877, p. 106].

Implicit in this definition is the idea of supply limitation that a channel will convey as much sediment as its capacity allows, unless there is insufficient material to be transported. More widely influential was Gilbert's seminal U.S. Geological Survey Professional Paper *The Transportation of Débris by Running Water*, published in 1914. It is this work that seems to provide the first formal definition of transport capacity in the English literature: "The maximum load a stream can carry is its capacity. … Capacity is a function of various conditions, such as slope and discharge, …. When a fully loaded stream undergoes some change of condition affecting its capacity, it becomes thereby overloaded or underloaded. If overloaded, it drops part of its load, making a deposit. If underloaded, it takes on more load, thereby eroding its bed" [*Gilbert*, 1914, p. 35].

He goes on to say that capacity is a function of discharge, slope, débris size, and channel form (channel depth/width) ratio. While the concept of transport capacity is asserted without support from further reference in the 1877 report, his statement that "the maximum particles which streams are able to move are proportioned to the sixth powers of their velocities" [*Gilbert*, 1877, p. 104] shows that he was at least aware of the work of *Leslie* [1823] or more likely *Hopkins* [1844] on entrainment by this time. In 1914, however, Gilbert suggests that the idea can be traced at least as far back as 1786 in the work of Dubuat, who demonstrated experimentally that the sediment discharge in a water flow was proportional to a square of excess velocity (flow velocity minus the flow velocity required for entrainment). These results were certainly still in use almost a century later, for example, in the study of *Lechalas* [1871]. Both of these authors were working in a tradition of applied hydraulics, interested in the navigation of rivers and the building of canals.

Although *Gilbert* [1877, 1914] provides the first uses of the term in the geomorphological/geophysical literature, it was largely ignored or misunderstood by this community until championed by Strahler's advocacy of a process-based geomorphology in his papers of the early 1950s [*Strahler*, 1950, 1952]. (See, for example, Holmes' review of 1915 in which he suggests that the wide variability in observed rates means that these sorts of observation are unlikely to be useful in dating geological deposits and thus unlikely to be of broader use. But, by contrast that "experiments of these kinds, while of considerable theoretical interest, serve only as a check to undue speculation" [*Holmes*, 1915, p. 134] and what is needed is more field study.) However, the engineering aspects of his work meant that Gilbert influenced the hydraulic engineering literature more directly, for example, being cited in the work of Hans Einstein [*Einstein*, 1942, p. 567], originally working in the Zürich school but then later for the U.S. Department of Agriculture [e.g., *Meyer-Peter et al.*, 1934; *Einstein*, 1941, 1942, 1950]—but it is interesting that the first two of these papers do not mention the term capacity; it is only once that Einstein was more fully embedded in the U.S. system that he started using the term in print. From the 1950s, the term transport capacity is firmly established in fluvial geomorphology.

Apparently separately, in the aeolian literature Bagnold started to develop a concept of "sand-saturated wind" [*Bagnold*, 1936]. This concept arose from his experimental work to determine how much sand a given wind (shear) velocity could entrain. From this point on, the concept of capacity was established in the aeolian literature. Furthermore, because of Bagnold's wider interests in both coastal geomorphology [*Bagnold*, 1946] and what he considered his "hobby" [*Bagnold*, 1990, p. 161]—fluvial geomorphology [*Bagnold*, 1966, 1973]—

as well as granular flows [*Bagnold*, 1956], the terminology gained wider purchase. Bagnold's work seems to be an example of an independent, parallel development of a concept, in that he did not seem to be fully aware of Gilbert's legacy until his own direct forays into hydraulic engineering in the 1950s (despite an earlier, practical start in this area [*Bagnold*, 1990, p. 9]).

In the hillslope domain, a further apparently independent concept of transport capacity is introduced by *Ellison* [1947] in his work on soil erosion by water, although it was not taken further until the highly influential paper of *Meyer and Wischmeier* [1969]. The latter clearly uses concepts from the fluvial literature but cites neither Gilbert nor Einstein nor Bagnold as a basis for a broader concept of transport capacity in soil-erosion studies. The first explicit discussion of a transport capacity in relation to débris flows does not come until the 1990s but also seems influenced by the fluvial literature. However, an example where the concept has been used much less is in glacial geomorphology. In the rare case where it has been used, there seems to be an influence from the fluvial domain [*Alley et al.*, 1997], and there needs to be an evaluation of the extent to which the concept is applicable to the complex glacial system.

Today, the concept of transport capacity influences most branches of geomorphology. It is widely presented in introductory texts, both in general terms: "the rate of transport is limited by the transport capacity of the process, which is defined at the maximum amount of material the process can carry" [*Holden*, 2008, p. 302] and in relation to specific processes, for example, in the fluvial literature "the transport capacity of the stream…can be viewed as being directly a function of flow discharge and slope" [*Robert*, 2003, p. 11], "[t]he rate of bedload transport is almost entirely a function of the transporting capacity of the flow" [*Robert*, 2003, p. 81], "capacity refers to the volume of material that can be removed for any given flow condition" [*Robert*, 2003, p. 146], "most bedload formulae aim to determine the rate of bedload transport as a function of the transport capacity of the flow. This is done by calculating the excess flow capacity above a critical threshold, at which transport is initiated" [*Charlton*, 2008, p. 110], "stream power determines the capacity of a given flow to transport sediment. This is the maximum volume of sediment that can be transported past a given point per unit time" [*Charlton*, 2008, p. 93], "sediment transport rate (capacity) … The sediment transport rate is the amount (weight, mass, or volume) of sediment that can be moved past a given width of flow in a given time" [*Bridge*, 2002, p. 60] and in relation to the hillslope domain "because soil creep moves the regolith material…transport is always at the transporting capacity" [*Holden*, 2008, p. 309].

However, even within these examples, it is clear that there are differences in the terminology used. Furthermore, the application of the term in the research literature does not seem to be consistent within or among different process domains. For example, some authors use it to denote the maximum possible rate of sediment transport by a particular flow [e.g., *Abrahams et al.* 2001; *Abrahams and Gao*, 2006; *Foster and Meyer*, 1972], while others use it to denote the potential for transport once grain resistance has been overcome [e.g., *Davies*, 1988a; *Eaton et al.*, 2006; *Ferguson*, 2003] and yet others as the transport rate under an equilibrium condition [e.g., *Bagnold*, 1936; *Celik and Rodi*, 1991; *Gilbert*, 1877; *Gomez and Church*, 1989; *Zhang et al.*, 2009]. This flexibility in terminology is problematic. Not only does it make communication difficult both within the discipline and also especially in interdisciplinary studies [*Bracken and Oughton*, 2006]. Furthermore, from an epistemic perspective, if transport capacity can refer to multiple real-world processes, then it becomes all but impossible to test the idea, and it appears that only ancillary hypotheses are being evaluated rather than the core concept itself.

Wainwright et al. [2008] also criticized the use of the term in hillslope studies from both theoretical and practical considerations and in particular the difficulties it poses when moving from one process domain to another. In the context of this review, they specifically note that erosion that shows a transition from concentrated overland flow erosion and débris flow erosion (as observed, for example, by *Oostwoud Wijdenes and Ergenzinger* [1998]) could never be predicted by a transport capacity-based model, as it would not allow sediment fluxes to become high enough to allow the transition to occur. They also demonstrated that both theoretical developments and the limited testing that had been carried out were limited to steady state conditions and were therefore unlikely to be representative of dynamic conditions.

In the following sections we provide an overview of how the concept has developed and been defined in the different process domains of fluvial, aeolian, coastal, hillslope, débris flow, and glacial geomorphology areas of the discipline of geomorphology. We evaluate the conceptual development especially in terms of an undercurrent of changes in approach to measurement and monitoring: not least the shift from deterministic

to stochastic underpinnings of process understanding; the dynamics in space and time and thus the move away from simple, steady state assumptions [*Wainwright et al.*, 2008]; and the need to look at fluid-sediment systems in an integrated way in line with broader developments in complexity theory. By adopting a realist perspective, we argue that advances in all of these areas suggest that existing approaches are limited both conceptually and practically. We conclude by suggesting how the discipline might move forward for providing unified concepts of sediment-transport modeling.

2. Sediment-Transport Capacity in Fluvial Sediment Transport

Gilbert's concept of transport capacity in rivers was defined based on the assumption that sediment-laden flow is in equilibrium, so that the amount of sediment carried into a stream cross section or reach is equivalent to that transported out of it. Consequently, neither deposition nor erosion occurs in this cross section or reach, and hence, the associated channel morphology remains stable [*Gilbert*, 1877, 1914]. The linking of the ideas of capacity and equilibrium has been used to interpret the shapes of stream profiles [e.g., *Mackin*, 1948] (but see discussions by *Leopold* [1980] and *Bracken and Wainwright* [2006]) and implied in regime theory, which is the concept that river channels adjust toward a dynamic equilibrium when it is just able to convey the water and sediment supplied from upstream [e.g., *Lacey*, 1930; *Knighton*, 1998; *Ferguson*, 2008]. The original idea of transport capacity was further developed by *Einstein* [1950] who considered that capacity occurred when the rate of sediment supply was equal to the transport rate and thus when the channel profile was in equilibrium. Most subsequent uses in the fluvial literature are based on one or both of Gilbert or Einstein.

However, the prevalent usage of sediment-transport capacity has been complicated for two main reasons. First, sediment in a river has been considered to be transported in different modes. Coarse particles travel through intermittent contact with the bed [*Graf*, 1971]. Thus, their movement is supported by the bed, and the associated load is called bedload. Finer particles often move in suspension because their settling velocities are more easily balanced by turbulence. The associated load is referred to as suspended load [*Vanoni*, 1975]. Secondly, sediment transported in natural rivers is heterogeneous in size. Thus, both modes of transport can occur for a single flow condition. A further consequence of mixed grain sizes is that sediment-transport rates are controlled not only by river hydraulics but also by the interaction between different sized particles within transported sediment and bed materials [*Wilcock*, 1998, as already recognized in *Gilbert*, 1877], and the rate of upstream sediment supply. Sediment-transport capacity has thus been evaluated to date in terms of sediment—either homogeneous or heterogeneous—transported by bedload and in suspension.

2.1. Transport Capacity for Bedload

Gilbert explicitly indicated in his definition of transport capacity that his flume experiments were related to bedload transport only [*Gilbert*, 1914, p. 35]. The concept of bedload-transport capacity has been used as a benchmark against which the ability of a stream to transport sediment can be measured and compared. Furthermore, it has been widely used to determine degradation and aggradation rates on river channel beds and to understand if bedload transport is primarily constrained by sediment supply (i.e., supply-limited rivers) or river flow hydraulics (i.e., transport-limited rivers) [*Gilbert*, 1877; *Andrew*, 1979; *Hicks and Gomez*, 2003; *Jackson and Beschta*, 1982; *Leopold et al.*, 1964; *Lisle*, 2008; *Mackin*, 1948; *Reid and Dunne*, 1996; *Sear*, 1996]. Quantitatively, the bedload-transport capacity has been regarded as the bedload-transport rate directly obtained from the flow in equilibrium, regardless of whether the flow is supply or transport limited [*Ferguson et al.*, 1989; *Gomez and Church*, 1989; *Wilcock et al.*, 2001] or more commonly inferred by comparing the measured transport rates with those predicted by bedload-transport equations [*Lisle and Church*, 2002; *Marti and Bezzola*, 2006; *Mueller and Pitlick*, 2005; *Reid et al.*, 1996; *Warburton and Davies*, 1998]. Determination of the bedload-transport capacity in these applications is fundamentally supported by the common assumption that bedload-transport capacity is the transport rate of a stream in equilibrium [*Gomez and Church*, 1989] that can be predicted by one of the established bedload-transport equations [*Graf*, 1971; *Hicks and Gomez*, 2003; *Mueller et al.*, 2005]. However, numerous studies have shown that no single bedload-transport equation is applicable to all natural gravel bed rivers [*Almedeij and Diplas*, 2003; *Barry et al.*, 2004; *Bathurst et al.*, 1987; *Einstein*, 1941, 1942; *Gomez and Church*, 1989; *Martin*, 2003; *Reid et al.*, 1996] (Figure 1). This observation suggests that for given hydraulic and channel conditions, different equations may give rise to different bedload-transport rates, which is logically confusing as the "maximum load" specified in the concept of transport capacity should be a unique value. However, even in Gilbert's first statement on the concept, he suggested
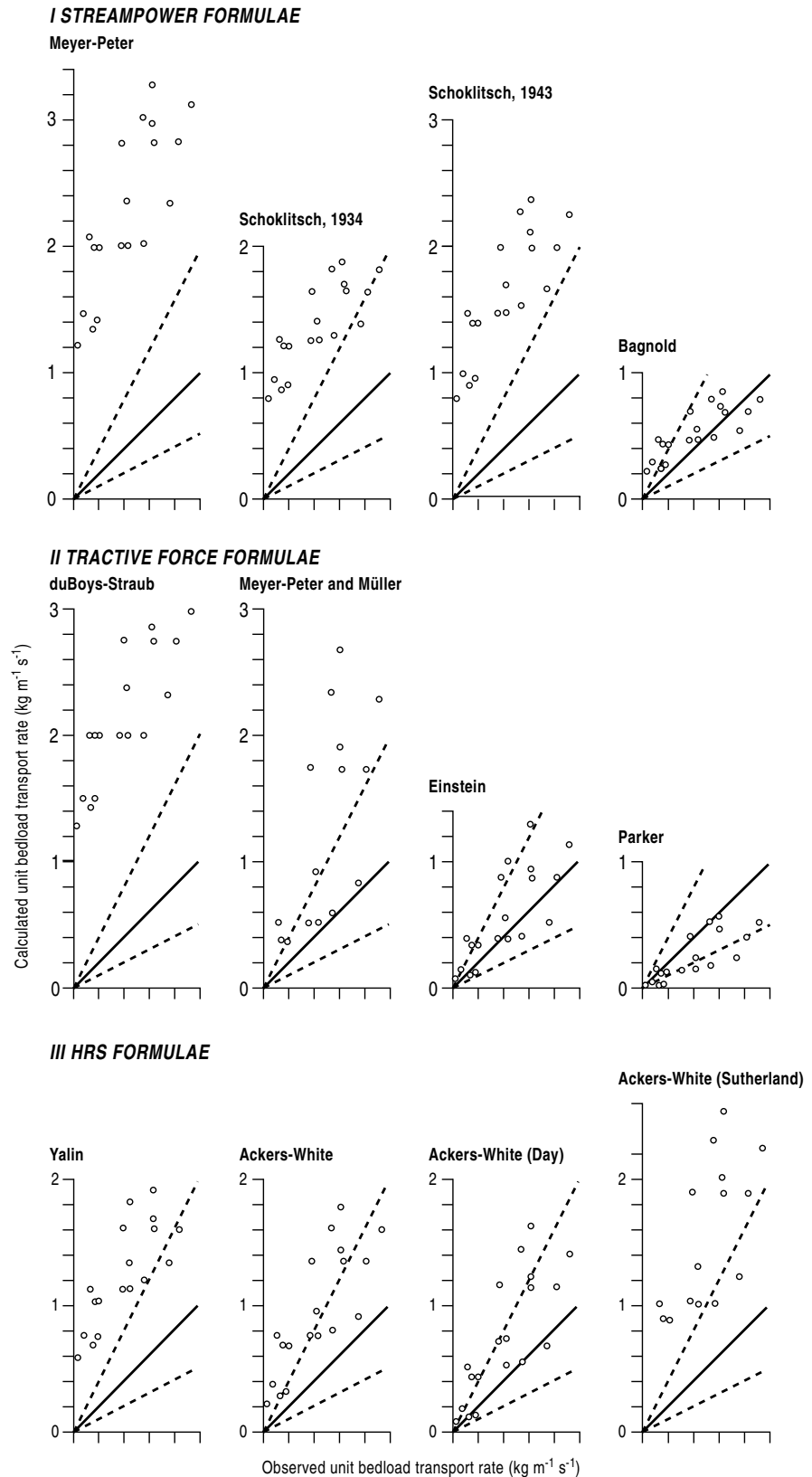
Figure 1. Comparisons between observed and calculated bedload transport in Elbow River, Alberta, Canada, data for different bedload formulae, illustrating that each formula produces contrasting estimates. (HRS: a group of formulae developed by researchers associated with the United Kingdom Hydraulics Research) [from *Gomez and Church*, 1989].

that "the capacity of a stream for transportation is greater for fine *débris* than for coarse" [*Gilbert*, 1877, p. 104], implying that the capacity is a function of particle size. Therefore, the concept of bedload-transport capacity needs to be revisited. Inasmuch as the behavior of bedload transport is ostensibly different between beds with homogeneous and mixed sizes, the concept will be reexamined and requantified for the two types of grain size distribution, respectively.

### 2.2. Bedload-Transport Capacity for Grains With Homogeneous Sizes

Bedload with homogeneous grains mainly occurs in flume experiments. In a flow under the equilibrium condition and over a movable bed consisting of grains of identical size, bedload-transport capacity has been considered equal to the bedload-transport rate of the flow, because it has been assumed that in such flows only one transport rate is available for a given bed shear stress and grain size. Consequently, this transport rate is the "maximum" rate that a given flow and sediment caliber can attain. This concept of transport capacity is consistent with Church's theoretical maximum transport rate that occurs over an unstructured bed composed of the same material [*Church*, 2006, p. 336]. However, the hydraulics of these flows is often complicated by the emergence of bed forms such as ripples, dunes, and pebble clusters wherein the shear stress responsible for the transport of bedload should be partitioned from the total shear stress [*Engelund and Hansen*, 1967; *Fredsoe*, 1993; *Gao*, 2012]. Although various correction methods for bed morphology have been developed [*Baldwin et al.*, 2002; *Engelund and Hansen*, 1967; *Stone and Murdoch*, 1989], selection of an appropriate method is not obvious in practice. To avoid such uncertainty, many experiments have been performed in flumes with flat, fixed beds [*Abbott and Francis*, 1977; *Francis*, 1973; *Hu and Hui*, 1996; *Niño and Garcia*, 1998; *Sekine and Kikkawa*, 1992]. However, by doing so, either these experiments do not test the underlying concept of capacity or they can only ever represent the behavior of sediment transport in very restricted conditions that seldom if ever occur in rivers. In the experiments of *Niño and Garcia* [1998], the saltation trajectories of individual bedload grains were measured using a high-speed video system, such that many dynamic properties of saltating grains (e.g., the saltation height, length, and mean grain velocity) could be measured. These measured variables, together with measured flow hydraulics and sediment sizes, have been subsequently used to solve governing equations established in terms of mass and momentum balance for individual bedload grains. This approach has led to various relationships between controlling variables such as shear stress and grain size (although issues relating to the interdependence of each of these variables are discussed below) and three dependent variables: δ_b the thickness of the bedload layer (m), C_b the mean volumetric concentration of bedload in this layer (kg kg^{-1}), and U_b the mean grain velocity (m s^{-1}) [*Hu and Hui*, 1996; *Krecic and Hanes*, 1996; *Lee et al.*, 2000; *Lee and Hsu*, 1994; *Sekine and Kikkawa*, 1992; *van Rijn*, 1984a; *Wiberg and Smith*, 1985]. By definition, the unit volumetric bedload-transport rate, q_b (m^3 m^{-1} s^{-1}), is then given by

$$q_b = \delta_b C_b U_b \tag{1}$$

Combining the relationships describing the controls on δ_b, C_b, and U_b in terms of equation (1) leads to various equations for bedload-transport rates.

However, many of the cited experiments that purport to determine transport capacity were performed by changing the sediment supply rate while keeping the flow rate unchanged, implying that a given bed shear stress and grain size may be related to several different bedload-transport rates. Therefore, bedload equations developed in this way are not capable of estimating a unique transport rate for homogeneous grains and are thus incompatible with the idea of a unique transport rate representing a capacity as suggested by this experimental approach. Again, either the underlying concept of transport capacity or the experiments used to try to demonstrate it have been found lacking.

The plane bed condition can also be achieved in flows over mobile beds [*Fernandez Luque and van Beek*, 1976; *Graf and Suszka*, 1987; *Niño et al.*, 1994; *Rickenmann*, 1991]. Bedload-transport rates measured in these flows have again been assumed to be at capacity. In these flows, bedload may be transported in either the saltation or the sheetflow regime [*Gao*, 2008]. The former involves flows with low and medium shear stress in which bedload moves by sliding, rolling, and saltating along the bed, while the latter contains flows with high shear stress where bedload is a loosely defined moving granular layer. In each regime, many bedload equations have been developed to predict bedload-transport capacities. For example, *Fernandez Luque and van Beek*'s [1976] and *Smart*'s [1983] equations are for capacities in the saltation regime, whereas *Wilson*'s [1966] and *Engelund*'s [1981] equations are for capacities in the sheetflow regime. However,
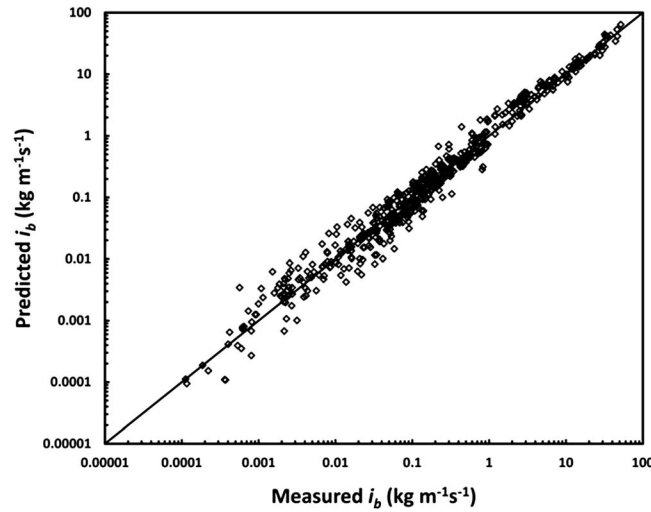
Figure 2. Measured submerged bedload-transport rate, i_b, versus predicted values using equation (3). The data were compiled from *Johnson* [1943], *Smart and Jäggi* [1983], *Gomez and Church* [1988], and *Recking* [2006] and those compiled by *Gao* [2003]. These data include all of the available experiments to date that transport bedload of homogeneous grains under the ideal condition and cover the full range of both the saltation and sheetflow regimes.

none of these can predict bedload in both regimes [*Wilson*, 1989]. Thus, if the transport rates predicted by these equations do conform to transport capacity, they do so only over the narrow range of hydraulic and sediment conditions under which they were developed and extreme caution should be employed when applying them more generally.

In an attempt to overcome this limitation, *Abrahams and Gao* [2006] compiled a large amount of bedload data with homogeneous grains that span a wide range of hydraulic and sediment conditions with mobile, flat beds. Using these data, they developed a general bedload equation aimed to predict bed load transport capacities in steady, uniform, and turbulent flows transporting bedload of homogeneous grains over flat, mobile beds, which is hereafter termed *the ideal condition*:

$$\varphi_b = \theta^{1.5} G^{3.4} \frac{u}{u_*} \qquad (2)$$

where $\varphi_b = \frac{q_b}{\sqrt{(\rho_s/\rho - 1)gD^3}}$ is the dimensionless bedload-transport rate [*Einstein*, 1950]

u is the mean flow velocity (m s^{-1});
$u_* = (g\,h\,S)^{0.5}$ is the shear velocity (m s^{-1});
g is the acceleration due to gravity (m s^{-2});
$G = 1 - \theta_c/\theta$;
$\theta = \rho hS/(\rho_s - \rho)D$ is the dimensionless bed shear stress (also known as the Shields parameter);
ρ_s and ρ are the density (kg m^{-3}) of bedload grains and water, respectively;
D is the median size (m) of bedload grains;
h is the mean flow depth (m);
S is the energy slope (m m^{-1});
θ_c is the critical value of θ for the initial movement of sediment.

The value of θ_c for each data set was estimated by plotting the volumetric bedload-transport rate q_b against θ, fitting a trend line to these data and extending it to the θ axis where $q_b = 0$ [*Abrahams and Gao*, 2006].

Gao [2012] introduced a new dimensionless bedload-transport rate $B = i_b/\omega$, where $i_b = q_b\,g\,(\rho_s - \rho)$ is the submerged bedload-transport rate (J s^{-1} m^{-2}), $\omega = \tau\,u$ is the stream power per unit area (or unit stream power) (W m^{-2}), and $\tau = \rho ghS$ is mean bed shear stress. He further demonstrated that equation (2) may be transformed into a much simpler form in terms of B:

$$B = G^{3.4} \qquad (3)$$

The predictive ability of equation (3) was evaluated using a set of bedload data not only independent of that used in *Abrahams and Gao* [2006] but also covering a wide range of hydraulic and sediment conditions. Combination of these two groups leads to an unusual new group of bedload data that represent almost the entire spectrum of the saltation and sheetflow regimes under the ideal condition. The good fit of equation (3) to these data (Figure 2) suggests that under the ideal condition only, bedload-transport capacity can be generally determined by a single equation. Inasmuch as G is quantitatively equivalent to $(P_g)^{0.5}$, where P_g is the relative frequency of grain-to-grain collisions during bedload transport at a given flow intensity [*Gao*, 2012], the general good performance of equation (3) may be attributed to the perspective of treating bed

a) Feed Flume

b) Recirculating, Flume
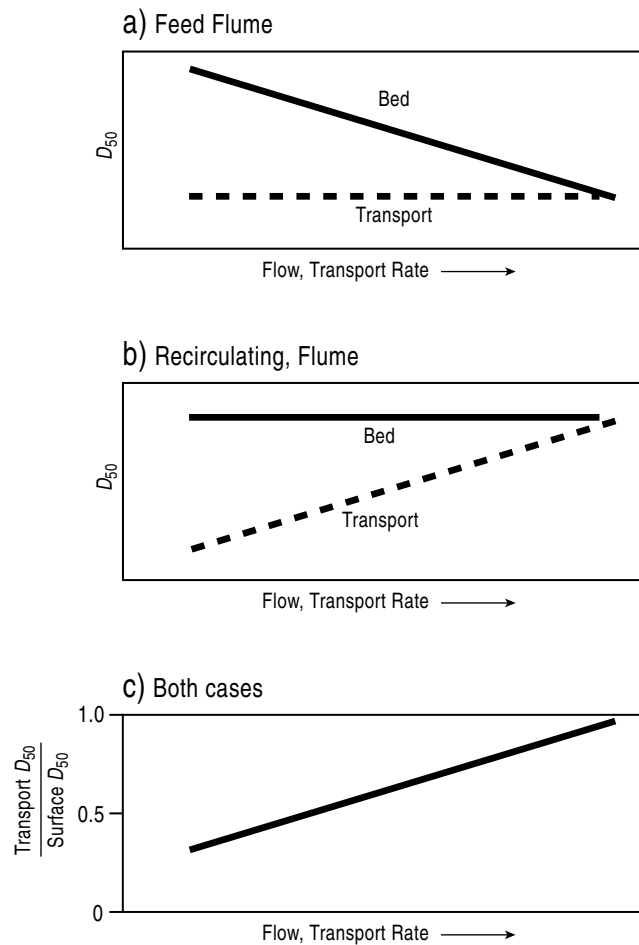
c) Both cases

Figure 3. Schematic illustration of grain size changes in the bed surface (D_{s50}) and transported (D_{50}) sediment. (a) A feed flume, only the bed surface changes. (b) A recirculating flume, the change is primarily in the transported sediment. (c) In both cases, transport coarsens relative to the bed surface. Thus, the same ratio of D_{50}/D_{s50} may be caused by (1) a low transport rate with small D_{50} or (2) a high transport rate with big D_{50} [after *Wilcock and DeTemple*, 2005].

load transport as a granular flow phenomenon [*Frey and Church*, 2011]. Although equation (3) only applies to flows under the ideal condition, it demonstrates the ability of *G* to characterize some aspects of the dynamics of bedload transport [*Gao*, 2012]. However, because the ideal condition rarely exists in natural rivers, it is of limited direct, predictive value. Furthermore, it could only ever be an incomplete component of a definition of capacity in more dynamic and variable settings.

2.3. Bedload-Transport Capacity for Grains With Heterogeneous Sizes

Bedload of heterogeneous grains is typically transported in gravel-bed rivers. A well-known phenomenon in gravel-bed rivers is bed armoring, in which the bed coarsens due to the preferential transport of fine bed material. This coarsening causes a reduction in transport rate and for fines to become less exposed to the flow (the hiding effect) [*Andrews and Parker*, 1987; *Egiazaroff*, 1965; *Einstein and Chien*, 1953; *Gomez*, 1983; *Lisle and Madej*, 1992; *Montgomery et al.*, 2000; *Sutherland*, 1987]. Therefore, for most of the time, only a fraction of grain sizes present on the bed of rivers is in motion [*Gomez*, 1995; *Lisle*, 1995], and so transport rates decrease compared to the values that have been assumed to represent theoretical capacity values as discussed above (Figure 3). During flooding, the armor layer may be broken

[*Ashworth and Ferguson*, 1989; *Clayton and Pitlick*, 2008; *Lisle and Smith*, 2003; *Parker and Klingeman*, 1982; *Wilcock and DeTemple*, 2005; *Wilcock and Southard*, 1989] and hence surface grains of all sizes are able to move as bedload at similar levels of bed shear stress, the condition similar to the strong form of equal mobility [*Parker and Toro-Escobar*, 2002]. Bedload transported under this condition is generally believed to be at capacity [*Gomez*, 2006; *Laronne et al.*, 1994; *Parker*, 2006; *Powell et al.*, 1999, 2001; *Wilcock and Crowe*, 2003], which has been taken to imply that bedload-transport capacity in gravel-bed rivers occurs only when the armor layer is broken and the transport rate is high, such as in peak flows during floods [*Lisle and Church*, 2002; *Wilcock and McArdell*, 1993].
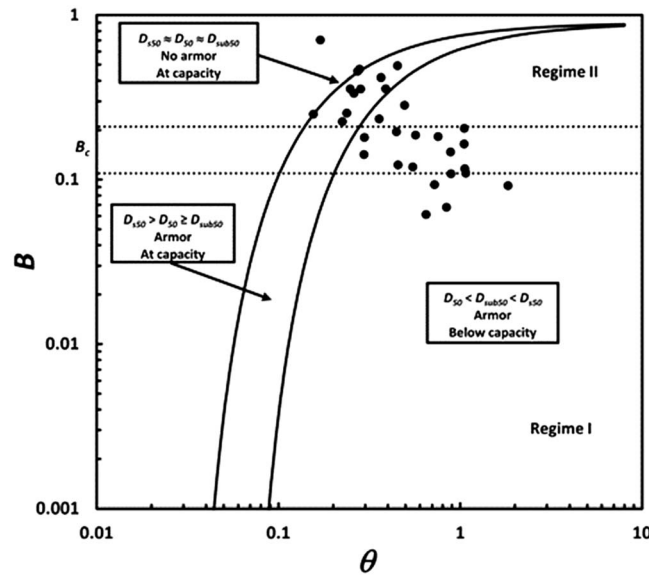
However, as stated above, bedload-transport capacity has also been regarded as the transport rate under the equilibrium condition, which may also happen during relatively low flow with an armor layer. These two concepts of bedload-transport capacity in gravel-bed rivers are inconsistent with each other if it is assumed that the flow alone controls the theoretical transport capacity. Furthermore, using the bedload data that were collected from a natural gravel-bed river [*Reid et al.*, 1995]—which have been widely accepted to represent transport-capacity flow [*Laronne et al.*, 1994; *Powell et al.*, 2001]—*Gao* [2011] demonstrated by comparing the measured bedload-transport rates with those predicted using equation (2) that the former is significantly lower than the latter, suggesting that even the relatively high transport rates in flows without the armor layer

Figure 4. The modified two-phase model. The two solid curves represent equation (5) with $\theta_c = 0.03$ and 0.06, respectively. The two dashed lines denote the boundary between the two regimes for the same two θ_c values. The areas between these curves and lines reflect the influence of the uncertainties in the determination of θ_c values. The dots are the bedload data reported in *Hayes* [1999] from a gravel-bed river significantly affected by a recent volcanic eruption (the data that have values of B greater than 1 are not included). Regime I is the area below the horizontal zone that includes two parts, the narrow area bounded by the two solid curves and the one on the right representing bed load transported at and below capacities, respectively. In the below-capacity area, bed load transport rate is relatively low for a given flow meaning the transport efficiency is relatively low and the median size of bed load D_{50} is small comparing to that of the bed surface, D_{s50} and substrate, D_{sub50}. In the at-capacity area, the transport rate is relatively high for the same flow suggesting that the relatively high transport efficiency and D_{50} is between D_{s50} and D_{sub50}. Regime II is the area above the dashed horizontal zone. It also has below-capacity and at-capacity areas. Flows in the former have a bed with an armor layer while in the latter do not. D_{50} in the former is relatively small while in the latter is equivalent to both D_{s50} and D_{sub50}.

are lower than the theoretical transport capacities of bedload with the equivalent homogeneous grains. Thus, the concept of bedload-transport capacity in gravel bed rivers (i.e., for heterogeneous grains) appears to have been applied differently from that for flows transporting bed load of grains with homogeneous sizes. Such an inconsistency reflects the problems with interpretation of the concept using a critical realist perspective.

In an attempt to avoid the inconsistency, bedload-transport capacity for heterogeneous grains was defined by *Gao* [2011] as the maximum possible transport rate a gravel-bed river can have for a given value of θ calculated using the median size of bedload grains D_{50} (m), although this approach may confuse issues further by providing yet another definition of what capacity might be. During high flows when the armor layer is broken, average grain-size distributions in bedload and the bed substrate are similar [*Andrews and Parker*, 1987; *Parker*, 2006; *Parker et al.*, 1982; *Parker and Toro-Escobar*, 2002]. When the channel bed is covered by an armor layer, bedload-transport rate has been related to the ratio of the median size of bedload grains (D_{50}) to the median bed-surface grain size (D_{s50} (m)) [*Parker*, 2006, p. 187, equations (3)–(48)]. Unfortunately, the relationship is uncertain due to variable sediment-supply rates, which adjust

both D_{50} and D_{s50} [*Buffington and Montgomery*, 1999a, 1999b; *Dietrich et al.*, 1989; *Gomez et al.*, 2001; *Lisle and Madej*, 1992; *Whiting and King*, 2003]. For example, the same ratio of D_{50}/D_{s50} may be caused by a low transport rate with small D_{50} or a high transport rate with a large D_{50} [*Wilcock and DeTemple*, 2005, Figure 3]. By examining a considerable body of bedload data obtained from natural gravel-bed rivers, *Gao* [2011] suggested that flows transporting bedload of heterogeneous grains at capacity should satisfy the following criterion:
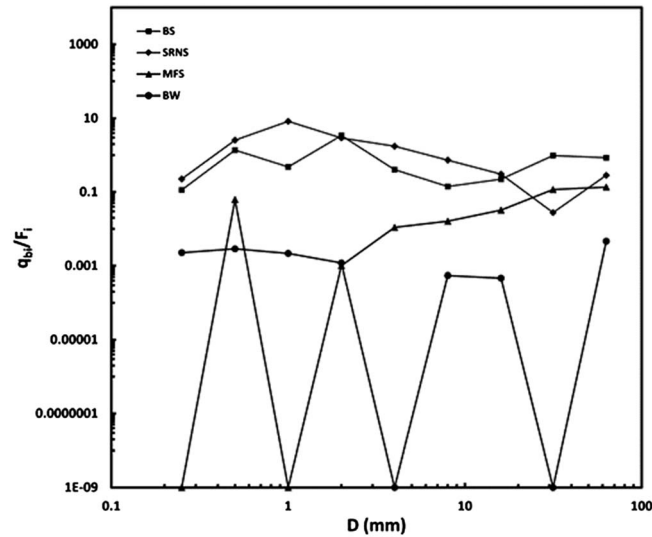
$$D_{50} \geq D_{sub50} \qquad (4)$$

where D_{sub50} is the median size of grains in the bed substrate (m). This criterion is compatible with the criterion above that the average grain-size distributions in bedload and the bed substrate are similar and is valid for flows both with and without the armor layer in gravel-bed rivers. Using bedload-transport data from flows of various hydraulic and sediment conditions, *Gao* [2011] then suggested that bedload-transport capacity for heterogeneous grains may be generally determined using

$$B = 0.9G^6 \qquad (5)$$

Equation (5) was further tested by showing its superior performance to the two well-known bedload-transport equations of *Meyer-Peter and Müller* [1948] and of *Bagnold* [1966] and its ability to identify correctly flows that had been assumed to be at capacity in six gravel-bed rivers with armor layers in Idaho [*Gao*, 2011]. In practice,

Figure 5. Plot of surface-based fractional transport rates (q_{bi}/f_i) against grain size fractions, D. Each curve represents a flow transporting bedload at capacity in one of the four gravel bed rivers in Idaho, USA.

θ_c may be either derived from transport measurements [*Church and Hassan*, 2002; *Mao et al.*, 2008] or estimated using a variety of statistical-, hydraulic-, or lichenometric-based approaches [*Gob et al.*, 2010; *Recking*, 2009; *Thompson and Croke*, 2008]. With a known value of θ_c, equation (5) represents a single curve in the plot of B against θ (the solid curve in Figure 4), which *Gao* [2011] suggests may be used to determine whether a given flow is transporting bedload at capacity (i.e., whether the flow is transporting bedload at a maximum rate). Comparing data representing four different flows measured in Simon River near Shoup, Idaho [*Boise Adjudication Team*, 2014], against the estimated maximum rate suggested that two achieved this rate and two did not (Figure 4). Further examining the ratio of the fractional transport rate q_{bi} to the surface grain size fraction, f_i, with respect to the associated median sizes of surface grains D_i showed (Figure 5) that flows with higher θ values tend to have higher fractional transport rates for coarser grains ($D_i > 10$ mm in Figure 5) and that the distributions of the ratios for "capacity" flows have similar patterns to those for "below-capacity" flows. Thus, flows that are considered at capacity may have similar bed-surface conditions to below-capacity flows. For the concept of transport capacity to hold one would expect the bed surface in a capacity flow to be finer in order to transport bedload at a higher rate. Therefore, the evidence suggests that the concept does not hold.

2.4. Significance of Bedload-Transport Capacities for Homogeneous and Mixed Grains

Gao [2011] suggests that bedload-transport capacities defined by equations (2) and (5) quantify the maximum loads a flow can transport for a given group of homogeneous and heterogeneous sediments. The definition is consistent with Gilbert's but more restrictive: "*bed-load transport capacity for heterogeneous grains is herein defined as the maximum possible transport rate a gravel-bed river can have for a given value of θ calculated using the median size of bed-load grains D_{50}. According to this definition, though a gravel-bed river with an armor layer may have several different available transport rates for a given θ value, it only has one maximum possible transport rate (i.e., the transport capacity)*" [*Gao*, 2011, p. 298] (emphasis original).

Since bedload in natural rivers is often transported at lower rates than these assumed capacity rates, as discussed in the previous section, these two equations (2) and (5) cannot be used directly for predicting bedload-transport rates. However, *Gao* [2011] suggested that they may serve as envelopes for all possible bedload-transport rates occurring in natural rivers and thus potentially have both theoretical and practical significance. In theoretical applications, by using the median grain size of the bed substrate (D_{sub50}), the dominant discharge, the representative channel slope, and the representative value of θ_c, the putative transport capacity calculated from equation (5) seeks to predict the average bedload-transport rate of a river in equilibrium. Therefore, *Gao* [2011] suggests that equation (5) can act as a simple extreme hypothesis [*Eaton et al.*, 2004; *Knighton*, 1998] to close the rational regime (i.e., which required to prevent erosion of the bed for design flow conditions) [see *Lacey*, 1930], albeit in very restricted conditions. In unsteady, nonuniform flow, equation (5) has been treated as a conceptual framework in comparison with the standard approach of Schumm to help interpret channel response to changing flow regimes qualitatively [*Phillips*, 2013], although no testing of the framework was carried out.

In practical applications, it has been suggested that equation (5) provides a quantitative benchmark for identifying the hydraulic conditions under which an armor layer begins to break [*Gao*, 2011]. It has long been observed that bedload-transport rate increases abruptly at certain values of water discharge (Q (m^3 s^{-1})) in

the plot of sediment-transport rate against Q. This increase has been attributed to additional sediment supply from the bed due to the breakup of the armor surface [*Emmett*, 1976; *Emmett and Wolman*, 2001] or the transition from selective transport (phase I) to equal-mobility transport (phase II) [*Jackson and Beschta*, 1982; *Lisle and Smith*, 2003]. However, the existing methods have generally focussed on determining the threshold values of Q, which are site specific [*Bathurst*, 2005; *Ryan et al.*, 2002; *Stoica and Sandgren*, 2006]. Equation (5) provides a general tool for identifying such hydraulic conditions. *Gao* [2011] argued that it can be assumed that bedload in flows without an armor layer is transported at capacity (a point to which we will return in section 8), so that the hydraulic condition under which the armor layer breaks θ_c must be a point on the curve in the plot of B against θ for a known value of θ_c. Furthermore, the breakup of an armor layer must result in accelerated increase of bedload-transport rates, which should be consistent with the inflection point along the curve representing equation (5). Mathematically, determination of the inflection point leads to [*Gao*, 2011]:

$$\theta = 3.5\theta_c. \tag{6}$$

When $\theta < 3.5\theta_c$, bedload-transport capacities occur with an armor layer; when $\theta > 3.5\theta_c$, bedload-transport capacities occur without an armor layer [*Gao*, 2011]. Based on these results, the two-phase model for identifying the breakup of an armor layer may be modified as shown in Figure 4. However, the asymmetry in observed transport rates as flows cross this threshold in increasing compared to decreasing flows [*Mao*, 2012; *Tunnicliffe et al.*, 2000] implies that there is again a problem with interpreting the flow as having capacity conditions at that single flow discharge. Besides leading to equation (6), equation (5) has also been used to estimate maximum possible sediment load for a potential extreme flood event as elaborated in *Gao* [2011], although this approach remains untested.

2.5. Transport Capacity for Suspended Load

Since particles moving in suspension tend to have smaller sizes than those moving as bedload, suspended load most commonly occurs in alluvial rivers with sand beds. The suspended-sediment-transport capacity of a given flow has been defined as "the maximum amount of sediment this flow can carry in suspension under equilibrium conditions for a particular sediment material" [*Celik and Rodi*, 1991, p. 192]. In such a capacity flow, sediment size in suspension is the same as that on the bed and the flow is in sedimentary equilibrium—that is, deposited particles can be easily replaced by eroded ones and "any further addition of sediments to the flow leads to a deposition of sediments on the channel bed without an increase of the suspended sediment concentration" [*Cellino and Graf*, 1999, p. 455].

The complication of determining suspended-sediment-transport capacity arises from the fact that the transport of suspended sediment is almost always associated with bedload-transport. For a given flow, distinguishing suspended sediment from bedload sediment is problematic and different criteria have been used to determine the two modes of transport [*Abrahams and Gao*, 2006; *Allen*, 2001; *Turowski et al.*, 2010; *van Rijn*, 1984b]. Thus, the two modes of transport may be better described as a continuum rather than as two distinct classes [*Parsons et al.*, 2015]. Therefore, it is often convenient to estimate the sum of suspended and bedload, which is conventionally referred to as the total load [*Graf*, 1971; *Hicks and Gomez*, 2003; *Julien*, 1998].

It has been assumed that these equations that developed theoretically (e.g., using *Bagnold* [1966], *Einstein* [1950], *Molinas and Wu* [2001], and *Yang* [2005]) predict suspended sediment (or/and total sediment) transport rates at capacity [*Graf*, 1971; *Hicks and Gomez*, 2003]. However, this assumption is not necessarily true because theoretically based equations often consist of several parameters that need to be determined using measured data. Unfortunately, in many equations, these parameters have been estimated using data obtained from flume experiments with fixed beds and variable sediment supply rates [*Galeone*, 1996; *Kronvang et al.*, 2012; *Kronvang and Bruhn*, 1996], which suggests that suspended sediment is not transported at a theoretical capacity [*Celik and Rodi*, 1991] (see also the discussion above). Testing the concept of capacity in this way is thus about testing the auxiliary hypotheses relating to parameterization, rather than the concept itself. For example, the vertical profile of sediment concentration has been quantified using theories of diffusion-convection and vertical exchange of particles [*Graf*, 1971; *Vanoni*, 1975; *Yalin*, 1972], which led to various equations that include the Rouse number, $R_o = v_\downarrow / \beta\kappa u_*$ where $v_\downarrow$ is the settling velocity of suspended sediment (m s^{-1}), κ is von Kármán's constant (—), and β is the ratio of the turbulent mixing

coefficient of sediment to the momentum exchange coefficient (−) [*Julien*, 1998]. The value of R_0 is primarily controlled by that of β, which is normally determined using measured profiles of sediment concentration in either flume experiments or natural rivers [*Julien*, 1998; *Simons and Şentürk*, 1992; *Cellino and Graf*, 1999; *Graf and Cellino*, 2002; *van Rijn*, 1984b].

3. The Concept of Transport Capacity in Aeolian Sediment Transport

In the context of aeolian sediment transport, the concept of transport capacity is most often expressed as being synonymous with the condition of steady state. Steady state implies that the flux of particles (e.g., sand in saltation) for a given wind shear is limited to an equilibrium value representing the saturation of the fluid flow with mobile particles. The assumption of steady state has been the foundation for developing models of sand transport by wind. A great deal of effort has been expended to define the theoretical relationship between the wind shear stress (τ_a, N m^{-2}) or wind shear velocity (u_{*a}, m s^{-1}, where $u_{*a} = \sqrt{\tau_a/\rho_a}$, and ρ_a is air density (kg m^{-3})) and the saturated flux of sand over a flat sand bed, typically denoted as q_{sat}, which has the dimensions mass per unit width per unit time [e.g., *Bagnold*, 1936, 1941; *Kawamura*, 1951; *Owen*, 1964; *Kind*, 1976; *Lettau and Lettau*, 1978; *Ungar and Haff*, 1987; *Sørensen*, 1991; *Sauermann et al.*, 2001]. *Bagnold* [1936] derived the transport rate of sand-saturated wind (q_{sat} (kg m^{-1} s^{-1})), which has come to be considered conceptually equivalent to transport capacity, as

$$q_{sat} = C_a \sqrt{\frac{d}{D_a}\frac{\rho}{g}} u_{*a}^3 \qquad (7)$$

where C_a is an empirical coefficient for sorting (apparently defined based on wind tunnel experiments: $C_a = 1.5$ for uniform sand, 1.8 for naturally graded sand, 2.8 for poorly sorted sand, and 3.5 for a pebbly surface), d is grain diameter (mm), D_a is the diameter of a standard grain (i.e., the one he first did his experiments on = 0.25 mm). The majority of theoretical models builds on the *Bagnold* [1941] expression:

$$q_{sat} = \frac{u_{*a}^3}{g}, \qquad (8)$$

although a few models have been developed that are based on mean wind speed u_a at a reference height z [e.g., *O'Brien and Rindlaub*, 1936; *Dong et al.*, 2011] as opposed to the shear velocity. In both cases it is necessary to include a threshold term [*Bagnold*, 1956], which sets a lower limit for transport as a function of either wind shear or mean wind speed.

In order to test the theoretical models, many experiments have been carried out using wind tunnels and atmospheric boundary layer flows to evaluate the predictive capability of models [e.g., *Chepil and Milne*, 1939; *Bagnold*, 1941; *Zingg*, 1953; *Williams*, 1964; *Sarre*, 1988; *White and Mounla*, 1991; *Greeley et al.*, 1996; *Butterfield*, 1999; *Namikas*, 2003; *Li et al.*, 2009]. The measurement of the sand flux in these experiments has been accomplished using traps of various designs and efficiencies as well as active sensors based on correlating flux with acoustic, piezoelectric, or optical signals, thus reflecting an evolution from time-averaged to point measurements. Total horizontal flux, q, is then calculated by integrating the vertical profile of point measurements of flux measured at specific heights [e.g., *Shao and Raupach*, 1992] or using traps with a continuous slot-like opening that extends to the approximate height of the saltation layer, thus integrating flux as a function of height during the measurement phase [e.g., *Gillies et al.*, 2006; *Dong et al.*, 2011].

Evaluations of how well measured horizontal flux rates compare with model-predicted values of saturated flux consistently show pronounced discrepancies [e.g., *Sherman et al.*, 1998]. The reasons for the discrepancies can be traced to difficulties in accurate measurement of the sand flux [*Ellis et al.*, 2009, 2012] and wind speed as well as in defining model coefficients [*Sherman and Li*, 2012], including coefficients in the "law of the wall," which provides the basis for determining wind shear velocity [*Bauer et al.*, 1992; *Sherman and Farrell*, 2008]. *Li et al.* [2010] observed variability in the von Kármán parameter κ during sand transport, which is treated as a constant in the log-law expression for boundary layer flow. One consistent observation from field studies is that the transport is highly variable in time [e.g., *Stout and Zobeck*, 1997; *Baas*, 2004; *Davidson-Arnott et al.*, 2009] and over space [e.g., *Bauer et al.*, 1996; *Gares et al.*, 1996; *Ellis et al.*, 2012], which suggests that transport capacity is not a time-independent property of the system (see discussion in section 8).

The degree of variability in wind flow and sand transport is much reduced in wind tunnels compared to the atmosphere, but wind tunnels create their own set of constraints on the transport system. The saltation process is altered by the dimensions of the wind tunnel, which introduce Froude-number effects [*White and Mounla*, 1991]. *Durán et al.* [2011] identify that a key difference between wind-tunnel and atmospheric boundary layer flows is the integral turbulent timescales, defined as the time difference beyond which velocities at a single place and two different times become uncorrelated. In wind tunnels the turbulent timescale coincides with the transport timescale ($\approx 1\,s$), while in a natural boundary layer with sand transport it is 10^3 times larger [*Durán et al.*, 2011].

With the observation that transport is typically unsteady in the atmospheric boundary layer, research has been directed toward an examination of this behavior in the field and in wind-tunnel experiments. *Stout and Zobeck* [1997] characterize the intermittency of saltation with a basic turbulence intensity parameter and a simple wind-strength index. *Stout* [2004] advances the use of intermittency to estimate the threshold wind speed based on measurements of saltation activity, mean wind speed, and the standard deviation of wind speed. The threshold is calculated during periods of saltation and reflects the conditions present at the time of measurement (e.g., grain-size distribution, surface moisture content, and relative humidity). The effect of unsteady wind conditions on flux was investigated in a wind-tunnel experiment by *Butterfield* [1998], who observed that transport was enhanced during introduced gusting periodicities between 6 s and 20 s, with rates in excess of those observed for steady winds of the same mean speed.

Spies et al. [2000] present a model developed from the work of *McEwan* [1991] that allows wind velocity fluctuations to be forced upon the flow. The numerical simulations of *Spies et al.* [2000] wherein they impose several different types of unsteady wind behavior to simulate the effect of a succession of gusts indicated that transport rate cannot respond to wind fluctuations greater than a frequency of 0.5 Hz. Their model also showed that the response time of q to increased changes in wind speed was on the order of 2 to 3 s, while following a decrease in speed, the response time of q was approximately 1 s longer, which they note is in agreement with experimental data [*Butterfield*, 1991; *Hardisty*, 1993]. Based on their modeling results, *Spies et al.* [2000] call in to question the legitimacy of modeling the transport system using the approach broadly defined by equation (8), as u_* is an average quantity of the turbulent boundary layer and cannot account for the effects of the fluctuating components of wind. They do note, however, that statistical properties of the flow such as the standard deviation and mean value of the fluctuations in wind speed scale with u_* and suggest that the influence of turbulence on the transport mechanism be pursued further.

Researchers examining the aeolian transport system have conducted experiments in the field and in wind tunnels looking for links between turbulence parameters and sediment-transport responses. In this alternative approach, q is hypothesized to be governed principally by turbulent fluctuations and semicoherent flow structures [e.g., *Baas and Sherman*, 2005; *Baas*, 2004; *Walker*, 2005; *Baas*, 2006] and does not invoke the concept of transport capacity. *Baas* [2006] carried out field measurements of wind and sand-transport activity at high frequency (20 Hz) spanwise to the flow and found that there was a complex interaction between turbulence and sand transport on three spatiotemporal timescales: (1) an external range on the order of 60 s, which represents longer-term transport conditions that scale with time-averaged wind characteristics (i.e., u_*); (2) the integral timescale and below, which represent different transport patterns that show dependence on wind speed (streamer families, nested streamers, and clouds with embedded streamers as identified by *Baas and Sherman* [2005]); and (3) the scale of individual streamers at times less than 1 s. According to *Baas and Sherman* [2005], these streamers are a visual representation of near-surface individual eddies that have translated down through the internal boundary layer, skim across the surface, and entrain/transport sand as they move downwind. The length scales of the streamers at their measurement location (i.e., width = 0.2 m and spanwise distribution = $0.9\,m^{-1}$) appeared to be stable and independent of wind speed (or wind shear). In this framework for analyzing sediment transport, boundary layer turbulence and the gust cycles will, to a large extent, control the transport. However, *Baas and Sherman* [2005] caution that these properties of the flow may differ greatly for different surfaces even if u_* were similar. Although research has demonstrated that turbulence and sediment transport are closely linked, there remains considerable challenge in developing transport models that are not based on mean flow and transport rates.

Aside from its use to identify maximal transport rates of sand, the concept of saturated flux, and the development of the system to reach an assumed saturation, has also been used in aeolian research to examine
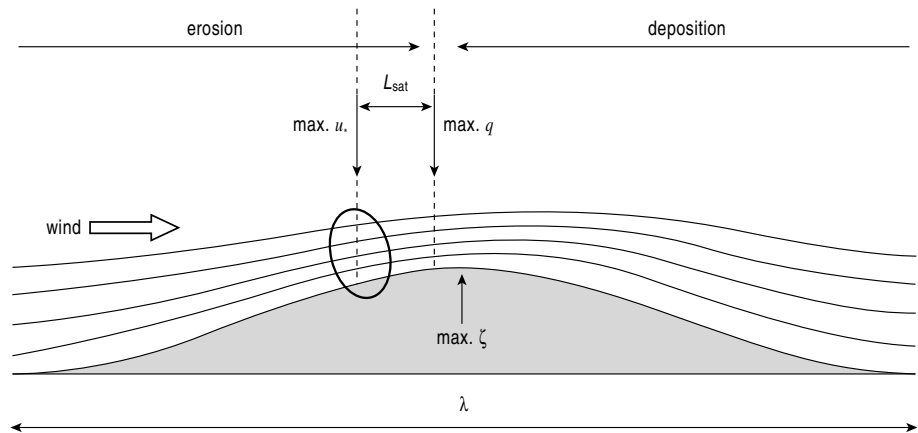
Figure 6. Schematic of the streamlines above a low-amplitude undulation of a sand surface in an aeolian setting. The maximum u_* is located at a distance upwind from the crest (maximum ζ) proportional to the wavelength λ. The sand flux maximum q_{sat} is located at a distance L_{sat} downwind, which separates the zones of erosion and deposition [after *Durán et al.*, 2011].

the time and length scales that the system manifests as it arcs from threshold to the attainment of saturated flux or the distance it takes to adjust to a new saturated state following a change in surface conditions (e.g., surface topography and roughness). Originally identified by *Shao and Raupach* [1992] as an overshoot effect, the saturation length scale has been linked to the scaling behavior of dunes affecting their profiles and wavelengths [*Sauermann et al.*, 2001; *Andreotti et al.*, 2002a, 2002b, 2010]; however, the hypothesized scaling dependency for the elementary size of dunes remains controversial [*Parteli et al.*, 2007a, 2007b; *Andreotti and Claudin*, 2007; *Andreotti et al.*, 2010].

Andreotti et al. [2002a] present a first-order relaxation differential equation to describe the behavior of these saturation scales:

$$T_{sat}\frac{\partial q}{\partial t} + L_{sat}\frac{\partial q}{\partial x} = q_{sat} - q_a \tag{9}$$

where t is time (s), x is horizontal distance (m), T_{sat} and L_{sat} are the saturation time (s) and length (m), respectively, and q_a is instantaneous sand flux ($kg\,m^{-1}\,s^{-1}$). T_{sat} defines the timescale at which a saltation cloud adjusts from one saturated state to another following an abrupt or sudden change in wind speed, which is quite fast (order of seconds). According to *Durán et al.* [2011], L_{sat} defines the length scale over which the flux reaches saturation as wind encounters an erodible surface, a change in roughness, or the adjustment from maximum u_{*a} to maximum q_a on the stoss side of a developing dune (Figure 6). *Andreotti et al.* [2010] argue that if grain inertia is the dominant dynamical mechanism limiting saturation, then L_{sat} should be independent of u_{*a} and scale as a function of the drag length (L_{drag} (m)), defined as the length needed for a grain to reach its asymptotic speed and scales as

$$L_{drag} = \frac{\rho_p}{\rho_f}\,d \tag{10}$$

where ρ_p is particle density ($kg\,m^{-3}$) and ρ_f is fluid density ($kg\,m^{-3}$). Using field data on sand-dune wavelengths and wind tunnel data, *Andreotti et al.* [2010] demonstrate that once rescaled, L_{sat} is around $2 \times L_{drag}$ within a 50% dispersion (Figure 7).

Hersen et al. [2002] offer a description of saturation length to conceptualize it and its role in sediment transport and bedform development. Briefly, as the entrainment threshold is reached and grains dislodge from the surface, they will accelerate toward matching the wind speed. During this acceleration phase the grain covers a distance depending upon its inertia in the transporting fluid, which scales as equation (10). As a grain descends and collides with the surface, it pushes some grains and splashes up others, and the splashed grains are, in turn, accelerated by the fluid flow, and as this process repeats, the sand flux increases until (assuming steady winds) grains cannot leave the surface except as a result of grains being deposited; i.e., the flux is saturated. The flux saturation length is proportional to this inertia length [*Sauermann et al.*, 2001] and will increase if the length of the saltating trajectories increases (scaling as defined by equation (11)).
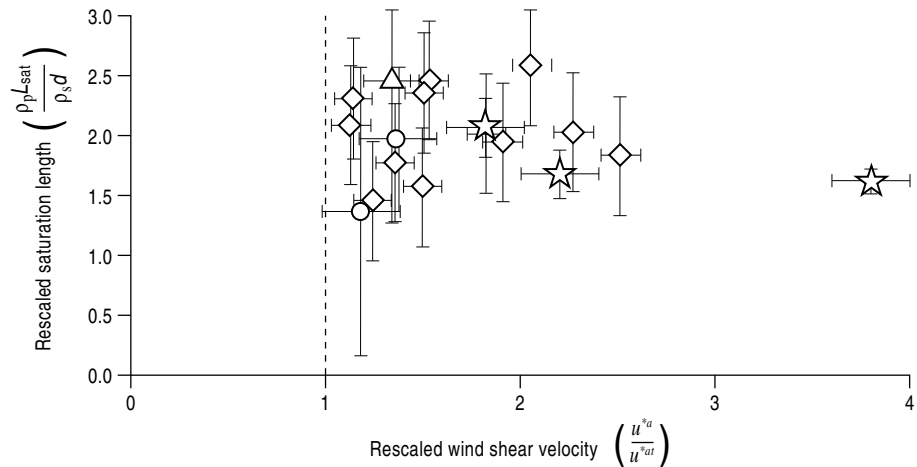
Figure 7. Saturation length L_{sat}, rescaled by the drag length ($L_{drag} = \frac{\rho_p}{\rho_f} d$), as a function of the wind shear velocity u_{*a}, rescaled by the threshold u_{*at}. Direct measurements, performed in a wind tunnel (diamond) and in the field (triangle), are compared to those determined from the initial dune wavelength (storms (star) and slipfaceless dunes (circle)) [after *Andreotti et al.*, 2010].

Elbelrhiti et al. [2005] observed that the length of protodunes in the Sahara increased linearly with mean grain diameter for the same wind regime, providing evidence for the control of L_{drag} on initial dune size. The theoretical prediction of the length at which dunes emerge from a flat sand bed requires linear stability analysis [e.g., *Andreotti et al.*, 2002a] that incorporates two components [*Andreotti et al.*, 2010]: (1) calculating the turbulent velocity field around an obstacle of low amplitude [*Fourrière et al.*, 2010] and (2) describing the sand transport around the saturated state, wherein L_{sat} is critically important. The outcome of this analysis gives the relationship between the emergent wavelength of the most unstable mode, L_{sat}, and the other parameters (e.g., u_{*a}, aerodynamic roughness in the presence of saltation, and z_0). *Durán et al.* [2011] note that the prediction of the emerging wavelength is essentially governed by L_{sat} and not sensitive to the flux relationship between q_a and u_{*a}.

The aeolian transport system extends in a continuum from particles creeping along the surface to those in reptation (movement along the bed without significant movement up into the airstream) and saltation finally through to dust-sized particles ($<70\,\mu m$ diameter [*Pye*, 1987]) carried in suspension. These particles are released from the surface by aerodynamic entrainment [e.g., *Kjelgaard et al.*, 2004a, 2004b; *Sweeney and Mason*, 2013], through ballistic impacts of saltating particles with the surface ejecting particles [*Shao*, 2000, 2004] or created by breakdown of soil aggregates [*Kok*, 2011]. These modes of transport are considered analogous to the bedload and suspended load for fluvial transport. The ability of the wind to transport the particles in suspension scales with the wind shear velocity (u_{*a}) balanced against gravitational settling velocity of the particle, which is a function of size and density. Particles will remain in suspension as long as the fluctuating vertical velocity component (w_a' (m s^{-1})) exceeds the settling velocity ($v_{\downarrow a}$ (m)), i.e.,

$$v_{\downarrow a} < \sqrt{\overline{w_a'^2}} \qquad (11)$$

and w_a' scales with u_{*a} [*Gillette*, 1977]. Due to limitations on the supply and delivery of dust-sized particles to the atmospheric boundary layer by constraints associated with surface conditions controlling the release of particles [*Gillies*, 2013], the transport of dust-sized particles will always be below capacity. During extreme dust-storm events, measured concentrations of particles $\leq 10\,\mu m$ aerodynamic diameter have exceeded 14,000 $\mu g\,m^{-3}$ (averaged over 24 h) at Mono Lake, CA [*Ono et al.*, 2011], which undoubtedly means that hourly concentration values were potentially much higher. *Orlovsky et al.* [2005] report that during dust storms in Turkmenistan visibility has been reduced to zero, which would require concentration of the suspended material to be much greater than any reported measurements. Even for light wind conditions, concentrations of dust ($<50\,\mu m$ geometric diameter) measured during haze events has exceeded 13,000 $\mu g\,m^{-3}$ [*Gillies et al.*, 1996]. The capacity of boundary-layer winds to transport dust-sized particles is likely never approached for terrestrial conditions.

The assumption of the aeolian sediment-transport system trying to or attaining saturation (i.e., capacity) has provided an important contribution to the explanation of transport rates, patterns, and bedform development [e.g., *Durán et al.*, 2011]. It is clear, however, that this framework cannot explain all aspects of the mechanism of sediment transport by wind. Recent advances in observing the relationships between turbulence characteristics in the flow and responses of the sediment to these forcings have provided insights into the mechanisms of the transport processes [e.g., *Baas*, 2006], without invoking transport-capacity concepts. The challenge, however, of the turbulence-controlled framework for sediment transport is in developing predictive models to quantify sediment transport using some parameterizations for the key turbulence properties that control the flux of the particles in transport. At this juncture in aeolian process geomorphology it would seem that both saturation and nonsaturation approaches still provide critical insights into understanding the sediment-transport system. It remains to be determined if these frameworks can be unified or even whether that is a necessary step to take.

4. Sediment-Transport Capacity in Coastal Geomorphology

Although the term sediment-transport capacity has not usually been explicitly employed in the field of coastal geomorphology, except in work directly by or influenced by Bagnold, the concept is implicit in the context of work on the rate of longshore sediment transport, which has been primarily developed from the standpoint of practical applications in coastal engineering. The history of the use of the term in coastal geomorphology mirrors the debate in the fluvial literature on the relationship of actual to theoretical values of transport capacity and raises issues of scales of observation/measurement in the applicability of the term.

Early work developed empirical relationships between driving wave power and longshore sand transport (see *Horikawa* [1978, chap. 5.4] for a summary). The major problem in Horikawa's analysis is the substantial variance over orders of magnitude in the parameters associating the relationship between measures of transportability and sand transported [see *Horikawa*, 1978, Figure 5.4.23; *Horikawa*, 1988, Table 4.2]. Such variation has to be seen in terms of the pragmatic attempts by investigators in establishing actual sediment transported as well as estimating the transport power involved at both prototype (i.e., full) and model scale. Most estimates of the relationship come from empirical statements of volume changes in profiles (i.e., some function of gross beach changes) versus wave measurements of often widely changing conditions and directions. The former, in particular, dealt with gross sediment changes and did not always differentiate between bed and suspended sediment loads (but assumed that median sand size was representative across the dynamic range) and had to make some assumptions as to limited or nil-offshore loss and dominant (if not total) longshore transport on actual beaches. It is also noticeable that initial measurements tended to derive from "infinitely" straight beaches, whereas most beach systems are not straight and as such induce subtle and scaled feedbacks in the variation of effective wave power especially after breaking where the strategic effects of morphodynamic relationships are dominant (see the discussion of *Short*'s [1999a, 1999b] work). Likewise, observations of wave power based on limited wave observations (often visual) or even from early forms of offshore wave recording provide very poor characterization of available work, leading to major variance between reality and observation. In particular, later differentiation between direct wave thrust power (at the breakpoint), direct longshore tidal current, and secondary current generation in the presence of breaking waves all combine to cause poor characterization by use of prebreaking wave parameters. Set against these difficulties of measurement, it is unsurprising that relationships between transport rate/capacity and wave power were difficult to obtain. Nonetheless, a body of literature did emerge (especially from the west coast of the U.S.) that supported a pragmatic positive relationship (at log scale) between wave forcing and sediment transported.

Typical of this pragmatic approach is that of *Caldwell* [1956], which is often cited as the initial exemplar work of relating transport volume to what has become known as wave power:

$$Q_l = 210[P \sin\alpha . \cos\alpha]^{0.8} \tag{12}$$

in which Q_l is the longshore volume transport of sand (originally measured in cubic yards d^{-1}), P is the incident wave power (millions of ft-lb/d/ft of shoreline), and α is the angle the breaking wave makes with the beach (°). Caldwell's work was trying to link a longshore sediment-transport rate (Q_l) to a generative process

power (P), though later investigators use P_l (defined as the longshore-directed component of P) where P_l is defined as

$$P_l = (E \, C_w \, n)_b \sin \alpha_b \cos \alpha_b \qquad (13)$$

where $(E \, C_w \, n)_b$ is the wave energy flux or power per unit wave crest length at the breaker point ($_b$) (as a power term, the units of $(ECn)_b$ are $\mathrm{W\,m^{-1}}$, though earlier definitions were in $\mathrm{dyn\,s^{-1}\,m^{-1}}$). Equation (13) is based on linear wave theory where (C_w) is the velocity of the waveform, the energy of the wave; (E) is the wave energy (proportional to the square of wave height); (n) a dimensionless calibration of waveform velocity (which increases from 0.5 in deep water to 1 in shallow water); and α_b is the angle between the breaker crest line and the shoreline. The sine and cosine terms translate the full onshore wave power to a directional thrust of wave power obliquely against the shoreline, i.e., the longshore component of wave power (P_l). This type of wave power derivation has become central as the driver of shoreline transport in subsequent analyses (see *Komar* [1999] for a general development of this approach) where a blend of observed versus theoretical statements of actual longshore transport have been developed to match against this type of statement of potential longshore power (i.e., transport capacity). A major difficulty has been specifying a highly variable wave power (fluctuating at seconds to hours) against sediment-transport rates that at the prototype scale tend to be net statements based on months to years.

Inman and Bagnold [1963] pointed out that equation (13) is both dimensionally incorrect and provides only an empirical relationship between observed wave power and sand transport, rather than what may be thought of as a more theoretical transport capacity: a similar confusion to that of the "potential-actual" debate of the fluvial dynamicists. Inman and Bagnold preferred to drop the Q_l measurement for what they term the "immersed-sediment weight transport rate" (I_t ($\mathrm{m^3\,L^{-1}}$)), which can be related to Q_l through

$$I_t = (\rho_s - \rho) g a' Q_l \qquad (14)$$

in which the use of both a' (the correction for pore space taken as 0.6) and the $(\rho_s - \rho)$ term means that only the immersed density of solid sediment transported is considered. The use of I_t—rather than Q_l—is based upon the sediment-transport theories of *Bagnold* [1963, 1966]. These authors proceeded to develop a theoretical underpinning for I_t to derive

$$I_t = K \frac{EC_w n}{u_0} \frac{\cos \alpha}{\tan \varphi} \overline{u_l} \qquad (15)$$

in which K is the ratio of the rate of work done in transporting sediment in relation to the total wave power available ($-$), $EC_w n$ is the rate of transport of energy of a wave ($\mathrm{W\,m^{-1}}$) as defined in equation (13), u_0 is the mean frictional velocity relative to the bed in the surf zone ($\mathrm{m\,s^{-1}}$), φ is the intergranular friction angle (°), and $\bar{u}_l$ is the mean longshore current velocity ($\mathrm{m\,s^{-1}}$). Inman and Bagnold further argued that if the position of rip currents can be assumed to be random in time and space along the beach, then a general expression for the rate of longshore sediment-transport capacity over a straight beach could be derived. They provide the rationale by which the immersed sediment weight transported by wave power can be related by

$$I_t = K \, P_l = K(EC_w n)\sin \alpha_b \cos \alpha_b. \qquad (16)$$

This relationship is plotted against field measurements of sand transport rates (Figure 8) and shows quite good agreement, yielding a value for K of 0.7 [*Komar*, 1999]. K henceforth became a significant concept in relating alongshore sediment transport to available wave power, underpinning both the original theoretical work by *Inman and Bagnold* [1963] and the empirical relations such as that compiled by *Horikawa* [1978] and *Komar* [1999]. Unless the nature and controls on K can be understood, any attempt to move beyond empirical predictions of longshore transport rate toward any process-based notion of longshore transport capacity will be hamstrung.

It has been suggested that K might vary with grain size [*Tanner*, 1979] and hence provide comparability with (limitations of) transport capacity formulae in aeolian and fluvial systems. However, there is little support for the notion that K decreases with D_{50} within the range of sand-sized sediments [*Komar*, 1980], though some (but questionable) support that such a relationship might exist if the size range is extended to include gravel and shingle [*Komar*, 1999]. However, data from gravel beaches are limited, but the increasing influence of particle shape with increasing size (within the gravel-cobble size range [*Orford and Anthony*, 2013]) means
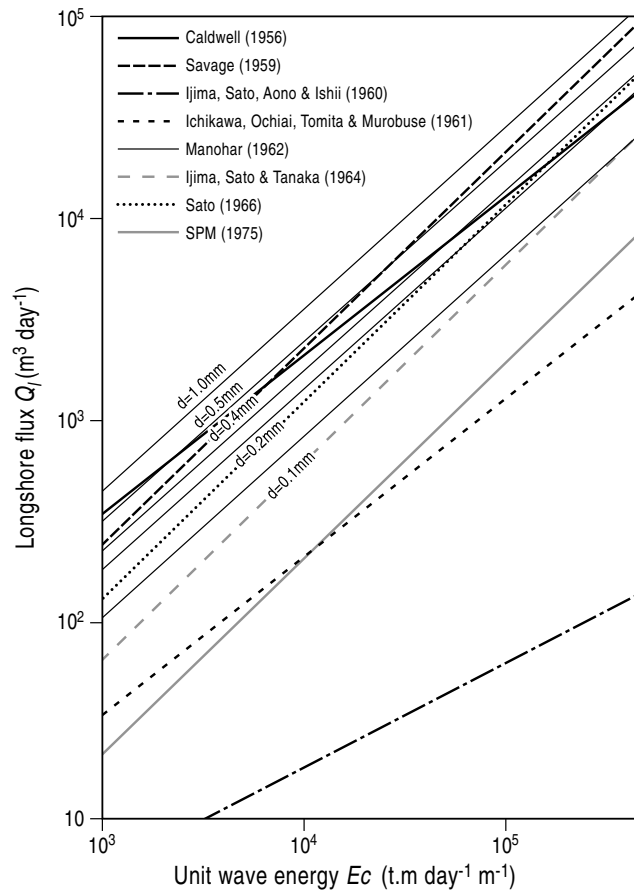
Figure 8. Different equations used to predict sediment transport under longshore conditions, showing the wide range of potential values for a specific wave energy [after *Komar*, 1999].

that a singular K value is unlikely. Other research has proposed that K depends on beach slope and wave steepness as well as grain size according to

$$K \propto S \left(\frac{H_b}{D_{50}} \right) \left(\frac{H_b}{L_\infty} \right) \qquad (17)$$

in which S is slope (m m^{-1}), H_b/L_∞ is a dimensionless measure of breaking wave steepness (breaking wave height (L) over offshore wavelength (L)), and H_b/D_{50} is the dimensionless scaled wave sediment size parameter (L/L). Furthermore, given that there is a positive relationship between sediment size and beach slope, the relationship will confound any attempt to find a simple relationship with grain size. Studies by *Saville* [1950], *Özhan* [1982], and *Kamphuis* [1991] have produced conflicting results on the relationship with wave steepness, but the suggestion is that K decreases with wave steepness (i.e., K reduces as short-period storm waves give way to long-period swell waves). Equation (17) also implies that transport rate increases with beach slope. However, given the advances in understanding beach morphodynamics [*Short*, 1999a], the above relationship also indicates that K is proportional to whether the beach system is reflective (steep and coarse sediment)

or dissipative (lowest slopes and smallest sediment). The implications for secondary current generation within a dissipative morphodynamic regime and the ensuing associated three-dimensional beach morphology changes with this energy domain underlie the difficulty of asserting a single K value across beach sediment size range. It also underlies the appropriateness of considering subsets of longshore transport capacity related to position in the tidal range and the ensuing potential of more consistent longshore beach morphologies where transport vectors both gross and net may be at base, unidirectional. Post–wave breaking on a reflective coarse clastic beach (subgravel size) may be the simplest context for K analysis. The likelihood of this context is low and only serves to indicate the difficulty of assessing the connection between actual and theoretical transport capacity.

The range and debate about the value of K may also reflect the possibility that the characterization of the dynamics of power to transport is still not adequately specified both in terms of actual parameters used and/or resolution of measurement of both sides of the power transport equation.

In recent years there have been attempts to resolve coastal sediment-transport rates with better determination of transported sediment volumes as well as clearer characterization and measurement of process variables. *Kamphuis* [1991] proposed a transport rate formula, the form of which was revised by *Mil-Homens et al.* [2013] as

$$Q_l = \frac{0.15\,\rho_s}{(\rho_s - \rho)}\, T_p^{0.89}\, (\tan\beta)^{0.86}\, d_{50}^{-0.69}\, H_{s,br}^{2.75}\, [\sin(2\,\theta_{br})]^{0.5} \qquad (18)$$

where Q_l is longshore sediment (dry mass in kg s^{-1}), $H_{s,br}$ is significant wave height at breaker line (m), θ_{br} is wave angle at breaker line (°), d_{50} is the median particle size in the surf zone (m), β is the beach slope (°), and

T_p = peak wave period (s). Note that the immersed weight has been changed to solid weight of sediment transported, while the explicit value of K has been lost, given the better specification of calculated driving power based on peak wave period and significant ($H_{67\%}$) breaking wave height. However, a dimensionless calibration element (0.15) is still required to constrain the predicted transport rates.

van Rijn [2014] provides the latest comprehensive analysis of existing transport equations and transport-rate data for both coastal sand and gravel systems. In this detailed examination of process he provides the following sediment-transport predictor equation for shorelines, based on the same approach as Kamphuis and Mils-Homens, but which has the ability to be applied across the sand to coarse gravel (0.1–100 mm) sediment range:

$$Q_l = 0.00018 \, K_{swell} \, \rho_s \, g^{0.5} \, (\tan \beta)^{0.4} \, d_{50}^{-0.6} \, H_{s;br}^{3.1} \sin (2 \, \theta_{br}) \tag{19}$$

As regular swell waves yield larger longshore transport rates than irregular wind waves of the same height (by a factor of 1.5), van Rijn proposed to take this effect into account by a swell correction factor (K_{swell}). This use of K is only loosely related to the earlier debate of K, as this K value is certainly not a constant and is designed to amplify the wave power potential of longer-period waves, rather than directly calibrate the potential versus actual wave power of the model. He identifies the value of K_{swell} based on the percentage of swell waves (p_{swell}) within the incident wave spectrum, such that as $K_{swell} = 1.05$ for $p_{swell} = 10\%$, $K_{swell} = 1.1$ for $p_{swell} = 20\%$, and $K_{swell} = 1.5$ for $p_{swell} = 100\%$. If swell is absent (or unknown), then $K_{swell} = 1$. The power of this predictor is in its highly reduced variance between predicted and observed transport across the sand and gravel size range [*van Rijn*, 2014, Figure 11] where about 80% of the 22 data points are within a factor of 2 of the 1:1 line. This is the current "state of art" for predicting longshore sediment transport and is a substantial improvement on the variance of the *Horikawa* [1988] statement.

All the approaches above are what can be termed computing a "total energy" capable of bulk transport. They apply a wave-energy parameter and angle of wave incidence to produce an "alongshore component" of wave power. Alternatively, there are a few studies that attempt to establish a modified shear stress/stream power formula based on the formulae that were originally used for estimating transport rates in rivers. *Morfett* [1990] explains the basic physics of this approach. Investigators have to change the flow field from a stream to that of a wave [e.g., *Bijker*, 1971; *Swart and Fleming*, 1980; *Bailard*, 1984]. This approach is an attempt to model the combined bed and suspended loads as a consequence of the distribution of transport rates in the surf zone and relating sediment transport to the rate at which energy is being dissipated per unit area, D. In an approach that draws directly upon the fluvial literature *Morfett* [1990] proposed an equation for longshore sediment transport based upon the notion of virtual wave power, drawing upon *McDowell*'s [1989] use of virtual stream power P_{str} (W m^{-2}) in which

$$P_{str} = \rho \left(u_*^3 - u_{*cr}^3 \right) \tag{20}$$

where u_* is shear velocity (m s^{-1}) and the subscript "cr" refers to the shear velocity at the threshold of sediment motion. The use of a third power on the shear velocity has indications of the continuity of Bagnold's approach in thinking of effective powers of shear stress from other sediment-transport regimes. *Morfett* [1990] then derived the equation

$$I_l = K \left[P(\sin \alpha)^{0.75} D_* \right] \tag{21}$$

where

$$P = P_+^{1.5} g^{-0.833} (\rho_s - \rho)^{-0.5} v^{0.167}. \tag{22}$$

where I_l is the transport rate (N s^{-1}), P_+ is the wave power that is dissipated in breaking and in overcoming friction (W m^{-2}), and v is kinematic viscosity (m^2 s^{-1}). D_* is a parameterized particle size (dimensionless), which Morfett developed from an empirical formulation based on the median grain size (D_{50}), relative density of immersed sediment, and water viscosity. He derived D_* as

$$D_* = D_{50} \left[g \left(\frac{\rho_s}{\rho} - 1 \right) v^2 \right]^{0.333} \tag{23}$$

Morfett reduced the overall equation to a power-law relationship (using log-transformed best fit linear regression) [*Morfett*, 1990, Figure 7] of the form

$$I_l = K \, P_* \, D_* \tag{24}$$

where P_* is P $(\sin\alpha)^{0.75}$. Once again, however, the equation depends on the constant K for which only an empirical determination based upon the range of observed transport rates is available. Morfett identifies K as 10.5×10^3 as a best fit regression constant. Hence, his final transport based on a dissipative wave energy model is

$$I_l = 10.5 \times 10^3 \, P_* \, D_* \tag{25}$$

Equation (25) attempts to provide a physical attempt at developing a transport-capacity statement for shorelines; however, its determination is dependent on the ability to develop a statement of energy dissipation at the shoreline which is somewhat intractable considering the easier measurement required for the bulk transport equations (e.g., equation (19)). This difficulty may explain why the literature has tended to expand on more defined and precise bulk-rate equations [e.g., *van Rijn*, 2014], rather than the energy-rate determinations as exemplified by Morfett.

5. Sediment-Transport Capacity in Hillslope Studies

The study of sediment transport on hillslopes comes mainly from attempts to develop process-based models to predict soil erosion. Early work was carried out by the U.S. Department of Agriculture Soil Conservation Service as a result of the Dust Bowl of the 1930s, resulting in the studies of *Zingg* [1940] and *Musgrave* [1947] that predicted sediment yield as a function of discharge, slope, and soil characteristics. The first explicit use of the concept of transport capacity seems to have been by *Ellison* [1947] in the first of his papers redefining soil-erosion studies. No citation is made of previous usage of the term, and its derivation is related to empirical observations made in one of Ellison's own experiments. *Meyer and Wischmeier* [1969] noted that Ellison's ideas had not been adopted, so little progress had been made in the intervening two decades in developing the empirical data that might underpin his approach. They consequently developed his conceptual model more fully (Figure 9) and made the first attempt to apply it mathematically. Using the relationship of *Laursen* [1958] from fluvial experiments, they estimated the transport capacity of flow as a function of flow velocity or discharge, and thus, the Meyer and Wischmeier paper seems to be the first instance when the fluvial literature is used to inform hillslope-erosion models. Although the idea of a transport capacity of rain is introduced by *Ellison* [1947] and then followed up by Meyer and Wischmeier, it has not subsequently been used, probably based on the assumption that measured rainsplash is always at capacity. The distinction between interrill and rill erosion, drawn by *Foster and Meyer* [1972] and *Meyer et al.* [1972, 1975], is crucial in underpinning the theory of transport capacity as applied to studies of erosion. As such, it underpins all but the most recent erosion models, although the distinction probably hides more of a continuum of behaviors [*Wainwright et al.*, 2008].

Transport capacity of interrill flows has been little investigated. *Moore and Burch* [1986] assumed that unit stream power as defined by *Yang* [1972] could be used to predict capacity in a continuum between interrill and rill flows, although most of their evaluations were on data sets relating to rilled slopes. They also demonstrated [*Moore and Burch*, 1987] that this form is conceptually equivalent to the standard form derived by *Julien and Simons* [1985] from dimensional analysis:

$$q_s = \zeta \, S^\lambda \, q^\mu \, i^\xi \tag{26}$$

where q_s is unit sediment discharge (assumed at capacity) (kg m^{-1} s^{-1});
q is unit flow discharge (m^2 s^{-1});
i is rainfall intensity (m s^{-1});
ζ, λ, μ, and ξ are empirical parameters.
where $\lambda = 1.3\,\psi$ or $1.375\,\psi$ and $\mu = 1 + 0.4\,\psi$ or $1 + 0.25\,\psi$, depending on conditions; ψ is the exponent in Yang's relationship between transport capacity and (excess) unit stream power. More recent empirical studies [*Everaert*, 1991] and overviews [*Prosser and Rustomji*, 2000] have also supported the use of equation (26), although most experimental approaches have assumed that measured sediment transport, q_s in equation (26), is equivalent to transport capacity without otherwise demonstrating the equivalence [e.g., *Everaert*, 1991].

The development of a process-based understanding of erosion and sediment transport by rillflow that is embedded in most process-based models of soil erosion of the later twentieth century has taken its lead from
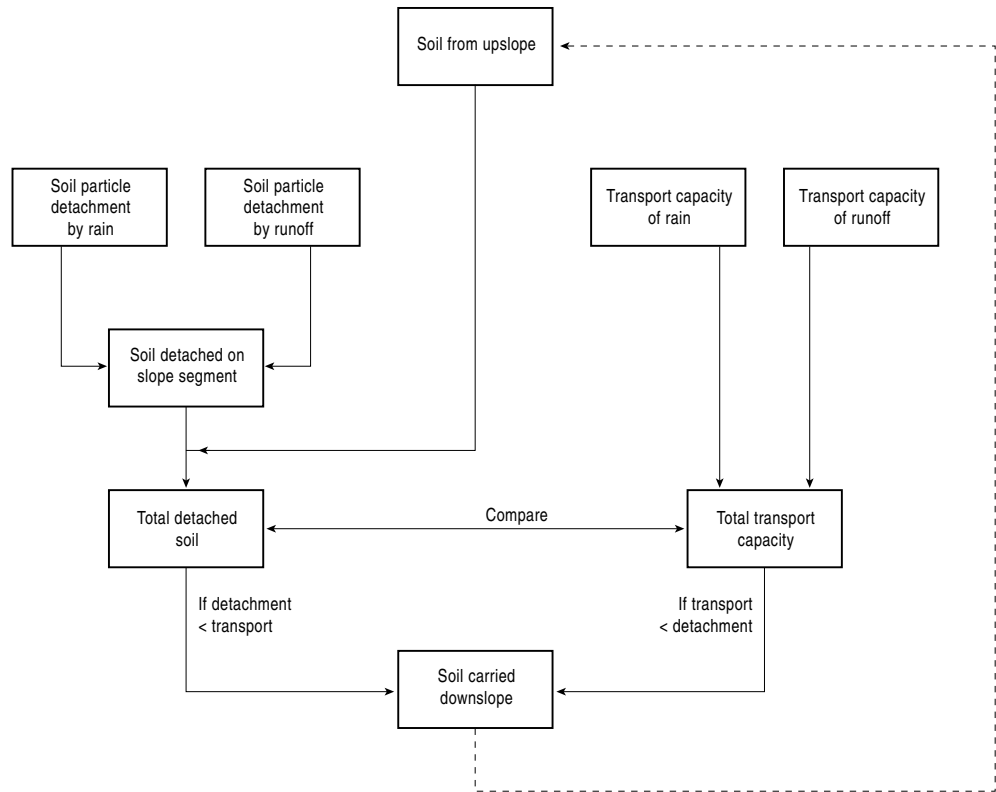
Figure 9. Conceptual model of soil erosion derived by *Meyer and Wischmeier* [1969] from *Ellison* [1947] and other sources.

the literature on river flow. The concept of transport capacity of rillflow is, therefore, an inherent component of these models. However, to transfer, largely empirical, equations developed for flow on very shallow gradients that is deep in relation to the size of most sediment transported to flow that is shallow and on typically much steeper gradients has proved problematic. *Foster and Meyer* [1972] developed a theory of erosion actually based on the assumption that under steady-state conditions:

$$\frac{D_F}{D_C} + \frac{G_F}{T_C} = 1 \tag{27}$$

where D_F is the net flow-detachment rate (kg m^{-2} s^{-1});
D_C is the detachment capacity of the flow (kg m^{-2} s^{-1});
G_F is the sediment load in the flow (kg m^{-1} s^{-1});
T_C is the transport capacity of the flow (kg m^{-1} s^{-1}).

By defining the detachment capacity as a linear relationship of transport capacity ($D_C = k\, T_C$), they produce the relationship

$$D_F = k\,(T_C - G_F) \tag{28}$$

It should be noted that this approach was arrived at by (a) a *qualitative* appreciation of laboratory and field observations and (b) an analogy that if transport capacity is a function of shear stress raised to a power (a simplified Yalin equation; see discussion below), then so too should detachment rate, with k in equation (28) simply the ratio of the coefficients in the detachment and transport-capacity relationships. At no point did they demonstrate that this analogy was supported empirically or theoretically. As early as *Bennett* [1974] pointed out that this argument was one that needed testing more fully, something that *Wainwright et al.* [2008] noted was yet to happen over 30 years later. Although some limited attempts have been made to verify the relationship [*Rice and Wilson*, 1990; *Cochrane and Flanagan*, 1996; *Merton et al.*, 2001], more
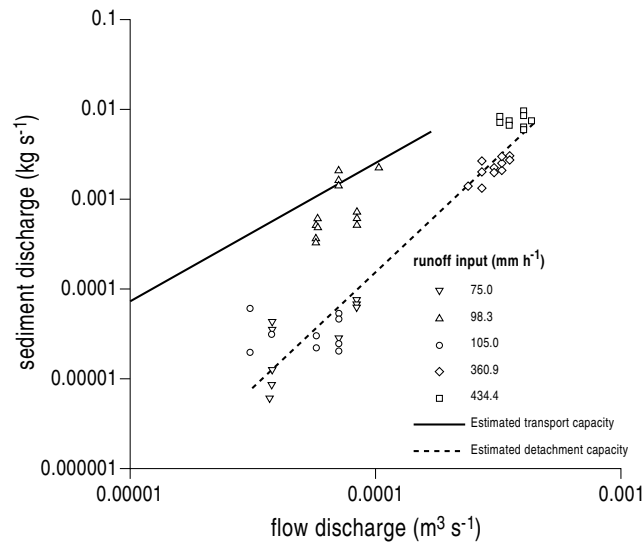
Figure 10. Comparison of experimental data with derived transport capacity and detachment capacity of overland flow on bare soil without sediment input at the top of the slope, but with flow addition at the rates specified [after *Schiettecatte et al.*, 2008].

detailed experiments by *Polyakov and Nearing* [2003] and *Schiettecatte et al.* [2008] demonstrate that the linear relationship in equation (28) does *not* hold (Figure 10).

In the Water Erosion Prediction Project (WEPP) model [*Nearing et al.*, 1989] soil detachment in rills is modeled as

$$D_F = D_c\left[1 - \frac{G_F}{T_c}\right] \qquad (29)$$

which is a straightforward rearrangement of the Foster and Meyer equation (equation (27)). Transport capacity is then determined from a simplified form of the Yalin equation, of the form

$$T_c = k_t\tau_f^{1.5} \qquad (30)$$

where τ_f (Pa) is the hydraulic shear acting on the soil and k_t (m$^{1/2}$ s^2 kg$^{1/2}$) is a transport coefficient.

Similarly, in the European Soil Erosion Model (EUROSEM) [*Quinton*, 1997; *Morgan et al.*, 1998]:

$$D_{FE} = k_f\,wv_\downarrow(T_C - C) \qquad (31)$$

where D_{FE} (m^3 s^{-1} m^{-1}) is the net detachment rate, k_f (dimensionless) is a flow detachment efficiency coefficient, w (m) is flow width, $v_\downarrow$ (m s^{-1}) is particle settling velocity, T_C (m^3 m^{-3}) is transport capacity, and C (m^3 m^{-3}) is sediment concentration. Transport capacity is determined from the work of *Govers* [1990] who found empirically that for particles between 50 and 150 μm, transport capacity could be expressed as

$$T_C = c(\omega_Y - \omega_{Ycr})^\eta \qquad (32)$$

where ω_Y (m s^{-1}) is Yang's unit stream power (defined as $u\,S$), ω_{Ycr} is critical Yang's unit stream power (= 4.0 mm s^{-1}), and c and η are experimentally derived coefficients that depend on particle size. The Limberg Soil Erosion Model [*de Roo et al.*, 1996] uses the same derivation, based on the work of *Govers* [1990]. GUEST (Griffith University Erosion System Template) bases its transport-capacity equation on Bagnold stream power [*Misra and Rose*, 1990; *Rose et al.*, 1998].

Both equations (29) and (31) share the assumption made by *Foster and Meyer* [1972] that detachment varies linearly with the difference between sediment in transport and transport capacity, though no published evidence supports this assumption [*Wainwright et al.*, 2008, p. 816]. Equation (29), taken from *Smith et al.* [1995], adds a modification to the detachment rate to take account of the fact that previously detached sediment will return to the bed, thereby reducing the sediment load. However, this modification is based upon the settling velocity of the particles and, therefore, assumes that all transported sediment is in suspension, which *Wainwright et al.* [2008] have demonstrated is unlikely to be the case for all but the finest sediments on most hillslopes under overland flow.

Equation (27) was proposed by *Foster and Meyer* [1972] and has empirical support in the work of *Alonso et al.* [1981]. However, later work by *Moore and Burch* [1986] and *Govers* [1992] shows the Yalin equation to perform relatively poorly in laboratory experiments of rillflow, and Govers further argues that all excess-shear formulae are unsuitable for rillflow. The form of equation (30) assumes that shear stress is well in excess of the critical shear stress to entrain particles. Inasmuch as the initiation of rills is defined by the point at which critical shear stress is exceeded, an equation for transport capacity that does not apply to this situation would seem inherently inappropriate. *Govers* [1992] further suggests that the apparent success with the Yalin formula achieved by *Alonso et al.* [1981] may lie in the restricted range of conditions tested, specifically that no gradient exceeded 0.07 m m^{-1}, which is scarcely representative of hillslope erosion.

Huang et al. [1999] carried out experiments where they could control the bed hydrology and found that identical flows where the slope had net drainage had lower estimated transport capacities than those with net seepage. *Sander et al.* [2007] used the Hairsine-Rose model to reinterpret these results as an emergence of an (effective) set of transport capacities under different conditions in that the Hairsine-Rose model does not define a transport capacity explicitly (as *Huang et al.* [1999] had also suggested for their results). They interpreted the results as reflecting the depositional layer protecting the bed once deposition starts to occur.

Polyakov and Nearing [2003] demonstrated that the Hairsine-Rose model would produce different values of transport capacity which they interpreted as hysteresis based on whether the value was reached from an excess or a deficit of sediment transport. They also noted that "transport capacity implies uniqueness" [*Polyakov and Nearing*, 2003, p. 42] and thus suggested that the above mentioned variable transport capacities for a similar flow and sediment condition suggest that these flows are not all at a putative capacity, the familiar situation discussed earlier in section 2.1 in alluvial rivers. Although one solution would be to use equation (3) or (5) to determine the unique transport capacity, it is important to recognize that neither of these approaches has been tested in hillslope settings. Some reasons fluvial approaches might not immediately transfer to the hillslope domain are elaborated in section 8.

Polyakov and Nearing also suggested that this mechanism might be an explanation for their results but proposed that they could also be due to changes in friction following deposition, protection of the bed by moving bedload, and differences in the energy required to start entrainment compared to continuing transport, as discussed in relation to aeolian sediment transport (and indeed, see *Ellison* [1947] on hillslopes). An alternative explanation of these patterns is the apparent emergence of multiple putative transport capacities as discussed in the section on bedload in rivers, and it is impossible to evaluate the appropriate explanation for hillslopes with the data available.

Equation (32) is derived from experimental work by *Govers* [1985, 1990, 1992]. However, in reporting this work, Govers noted that no equation that derives from the fluvial literature performed well over the full range of conditions that he tested and that significant gaps in the empirical base remained. Of note is the fact that Govers found the exponent η in equation (26) to be positive, a result also observed for interrill flow by *Everaert* [1991] implying that if critical stream power is fixed, then transport capacity increases with particle size, which is not intuitively obvious (and contradicts *Gilbert* [1877] as well as the later fluvial literature discussed above). Again, although further work is required to evaluate whether equation (3) or (5) could be used to overcome the limitations identified by Govers, there are limitations in relation to relative depth of flow that suggests such work would be fruitless (see section 8).

The significance of particle size for transport capacity is addressed poorly for rillflow. *Julien* [1987] suggested that estimates of transport capacity will be highly sensitive to particle size. The problem can be well expressed in a statement taken from *Govers* [1990, p. 45]: "Flow incision … will only occur when the transport capacity of the flow is sufficiently high to evacuate all the material that is transported into the flow path from the interrill areas." Since detachment in interrill areas is highly size selective and largely controlled by rainfall characteristics, its relevance to flow incision must be highly specific to particular soil characteristics. *Ferro* [1998] reinterpreted the grain-size dependence in Govers' and others' data by means of the Shields parameter and went on to suggest that the value should also be a function of flow depth relative to bed particle size (see discussion on relative roughness below). *Beuselinck et al.* [1999, 2002] suggested that the grain-size dependency increased above a threshold value and was not clearly distinguishable at lower flows. More recently, *Zhang et al.* [2011] estimated that transport capacity was proportional to $d_{50}^{-0.345}$ for a series of laboratory experiments on steep slopes (8.7–42.3%) but without the presence of rainfall.

6. Sediment-Transport Capacity in Débris Flows

The transport-capacity concept as developed for fluvial applications is not stated explicitly in most débris flow studies. The term "transport capacity" is often used without clear reference to established definitions or seminal papers in other literatures and is sometimes used to identify the overall volume of sediment (magnitude) of débris flows. The first explicit mention of "sediment-transport capacity" in the case of a débris flow is given by *Rickenmann* [1990, 1991]. He studied hyperconcentrated flows and debris flows using a recirculating system in a steep flume and explored the effect of an increasing fluid density and viscosity on the flow

behavior and the sediment-transport capacity. Even if hyperconcentrated flows are usually considered two phase, non-Newtonian flows, using a conventional (Newtonian) approach, he showed that an increased fine sediment concentration and flow density produced an increased coarse transport rate. However, at the highest suspended sediment concentrations, a decrease in the bedload-transport rates was observed and related to macroviscous effects. In a recent paper, *Shu and Fei* [2008, p. 973] explicitly state that the sediment-transport capacity for a hyperconcentrated flow is "the total amount of sediments that water flow can carry" and developed a formula for sediment-transport capacity of hyperconcentrated flow, in part based on *Bagnold*'s [1966] sediment-transport efficiency model for suspended transport. Based on the equilibrium equation of turbulent kinetic energy for solid and liquid two-phase flow, they obtained a structural formula for sediment-transport capacity, which includes the flow viscosity and flowing resistance coefficient for hyperconcentrated flows:

$$S_{v*} = p \left[\frac{f(\mu_r)}{\kappa^2} \left(\frac{f_m}{8} \right)^{1.5} \frac{\gamma_m}{\gamma_s - \gamma_m} \frac{U^3}{g R v_{\downarrow}} \right]^N \tag{33}$$

where S_{v*} is the sediment concentration at capacity ($m^3\,m^{-3}$);
p is an empirically determined coefficient ($-$);
μ_r is fluid viscosity (Pa s);
f_m is a coefficient of resistance ($-$);
γ_m is the specific gravity of the flow ($kg\,m^{-1}\,s^{-2}$);
γ_s is the specific gravity of the sediment ($kg\,m^{-1}\,s^{-2}$);
U is vertical mean flow velocity ($m\,s^{-1}$);
R is hydraulic radius (m);
N is an empirically defined coefficient ($-$).

Note that despite the theoretical derivation of this equation, it still contains two terms—p and N—and a function—$f(\mu_r)$—that need to be derived empirically is heavily dependent on analyses of standard fluvial flows and assumes that the specific gravity of the flow is unchanging as sediment transport increases.

It should be stressed that there are no widely accepted criteria that unequivocally differentiate between hyperconcentrated and débris flows (and by extension between hyperconcentrated and other water flows). Sediment concentration has been used to provide a rough distinction between the phenomena, and débris flows are usually considered to contain more than half their particles coarser than sand [*Pierson and Costa*, 1987]. However, débris flows are pulsating events, often anticipated by liquid surges and followed by mud surges, which lead to rapid variation of coarse sediment concentration, velocity, and viscosity [e.g., *Iverson*, 1997; *Marchi et al.*, 2002]. Generally, in natural streams débris flows tend to occur when slope is higher than $0.2\,m\,m^{-1}$ [*Takahashi*, 1991], but at these same slopes bedload and hyperconcentrated flows can occur as well. By analyzing débris flow, hyperconcentrated flow and bedload-transport events that occurred in Switzerland in 2005, *Rickenmann and Koschni* [2010] found a smooth, increasing trend between processes, if the transported sediment volumes normalized by the effective runoff volume were plotted versus the channel slope. This smooth transition between transport processes seems also supported by comparative application of bedload- and débris flow-derived formulae applied to calculate sediment concentration as a function of channel slope [*Rickenmann*, 2012]. *Prancevic et al.* [2014] suggest that this continuum can be understood in terms of a continuum of behaviors as Shields stress and slope angle vary, although more empirical work is required to develop their initial, experimental results.

The focus on débris flow volume is well justified because its quantification is of primary importance for hazard mapping. In field studies débris flow volume can be roughly estimated from postevent field survey or accurately quantified using LiDAR data [*Scheidl et al.*, 2008]. Débris flow volume is required for developing empirical relationships between planimetric deposition area and débris-flow volume and ultimately for relating geomorphological characteristics of fans to potential runout areas [*Iverson et al.*, 1998; *Berti and Simoni*, 2007; *Scheidl and Rickenmann*, 2010]. Furthermore, the volume is needed for assessing precisely the débris flow volumetric sediment concentration (C_V; volume of sediment divided by volume of water and sediment), which is required as input parameter in widely used débris-flow- and hyperconcentrated-flow-routing models [e.g., *Hsu et al.*, 2010]. Because in practical applications débris-flow volume is generally the unknown value,

the volumetric sediment concentration needs to be estimated as the ratio between the equilibrium concentration (C_D) and the volume concentration of solid fraction on the bed (C^*), for which the porosity of sediment in the bed can be used as a proxy. Expressed in terms of discharges, the débris-flow discharge at equilibrium conditions (Q_{DF} ($m^3 s^{-1}$)) could be calculated as

$$Q_{DF} = Q\frac{C^*}{C^* - C_D} \tag{34}$$

where C_D is the equilibrium concentration ($kg\,kg^{-1}$) and C^* is the volume concentration of solid fraction on the bed ($kg\,kg^{-1}$).

The equilibrium concentration (C_D) depends on the approach used to describe the rheological nature of the débris flow. It has been defined by *Takahashi* [1978, 1980], who suggested that the nature of débris flows is dominated by the collision between the coarse particles rather than the interstitial fluid. Takahashi's inertial-type approach is based on *Bagnold*'s [1954] dispersive regime and allows the velocity profile to be calculated, as well as the equilibrium grain concentration and the depth of a quasi steady state débris flow. Takahashi defines the equilibrium concentration as the concentration of a débris flow that produces neither erosion of the bed nor deposition onto the bed. Assuming a uniform distribution of sediments over the whole depth, the equilibrium concentration (C_D) is ultimately a function of the bed slope and can be calculated as

$$C_D = \frac{\rho_f\ S}{(\rho_s - \rho_f)\ (\tan\varphi\ -\ S)} \tag{35}$$

in which φ is the collision angle of grains which, according to *Bagnold* [1966], can be approximated with the internal friction angle of the sediments.

Even if in the model of *Takahashi* [1978] the role of interstitial fluid can be neglected, later studies suggested that the macroviscous flow regime should be taken into consideration [e.g., *Davies*, 1988b] and that cohesion and variable concentration and pressure over the flow depth should also be considered [*Chen*, 1988]. Attempts have been made to investigate intergranular stresses using a Coulomb mixture model, which accounts explicitly for stresses and interactions of distinct solid and fluid constituents and eliminates the need to specify rheologies of complex, multiphase moisture [*Iverson and Denlinger*, 1987]. Furthermore, from a series of experiments where liquid and solids were recirculated over a mobile bed, *Armanini et al.* [2008] showed an unstable stratification of collisional and frictional regimes across the flow depth, with implications for grain concentration and thus transport capacity. As to the calculation of equilibrium sediment concentrations, another alternative and promising theoretical approach based on the maximum entropy principle has been proposed by *Lien and Tsai* [2003]. Overall, many investigators have modeled débris flows specifying different rheological rules that govern flow behavior, and each model determines the sediment concentration at equilibrium.

In general, débris-flow routing can be simulated solving a momentum-conservation equation for the mixture of solids and fluid, mass conservation equations for the liquid phase and solid phase, and an equation for bed elevation change. Irrespective of the rheological approach to define the debris flow, which has to be reflected in the momentum-conservation equation [*Takahashi*, 2009], the débris-flow volume increases with bed erosion or decreases with deposition as calculated by the mass conservation equations. *Takahashi et al.* [1987] implemented formulae predicting erosion or deposition depending on the deficit or excess of sediment concentration in the débris flow with respect to the equilibrium volume as calculated by

$$\frac{E}{v_D} = \alpha_D\ \frac{C_D - C}{C^* - C_D}\frac{h}{d}\ (E \geq 0)$$
$$\frac{E}{v_D} = \beta_D\ (C_D - C)\frac{h}{d}\ (E < 0) \tag{36}$$

in which E is the erosion (positive)/deposition (negative values) rate ($m\,s^{-1}$), v_D is the average velocity of the débris flow ($m\,s^{-1}$), and α_D and β_D are experimental coefficients. Although *Takahashi et al.* [1987] describe some of their approach as having been derived by analogy with the bedload literature, they do not cite any of it explicitly. In the same way as the derivation of equation (35) is from theory (as with the analogous equation (27) developed for soil erosion by *Foster and Meyer* [1972], but which is not directly cited by *Takahashi et al.*

[1987]), there seems to have been little empirical testing of the model, especially its dependence on the definition of empirical coefficients.

Subsequent numerical models added to this approach, introducing, for instance, the effects of bed sediment size on bed erosion rate [*Egashira et al.*, 2001]. As shown by *Papa et al.* [2004], the vertical erosion rate is inversely proportional to the grain size of the channel bed. Also, the critical condition for the entrainment of a grain from the bed depends on the difference between the total shear stress and the yield stress acting on the bed surface and depends on sediment concentration of the débris flow. Despite these flume-based insights on drag forces needed to entrain sediments from the channel bed, sediments are also destabilized by macroscale processes due to undrained loading, impact loading, liquefaction, and streambank undercutting [*Hungr et al.*, 2005], and very little is known about the shear strength of bed sediments, bed stratigraphy, and pore pressure due to saturation. To overcome these mechanistic limitations, the magnitude of a débris flow is often predicted using empirical algorithms based on the geometry of the routing colluvial channel [e.g., *Ikeya*, 1981; *Hungr et al.*, 1984] or the size of landslides-prone areas upstream. An alternative approach, which also allows the association of a frequency to the magnitude of the expected event, is based on the analysis of rainfall thresholds able to trigger débris flows [*Jakob and Weatherly*, 2003].

It should be noted that all numerical and flume-based studies on débris flows and most of field empirical observations assume unlimited sediment supply and thus that equilibrium concentration (and thus a steady-state approximation to capacity analogous to that used in some of the fluvial literature) is always supposed to be reached. This assumption might appear reasonable in risk-based studies because hazard mapping or engineering countermeasures are better based on worst-case scenarios. However, even if such powerful events are likely to convey as much sediment as they can, relatively few geomorphological studies have questioned the supply-unlimited assumption. *Bovis and Jakob* [1999] stated that there are weathering-limited basins that require a certain "recharge period" prior to each débris flow event and exhibit a lower frequency of débris flow activity if compared to transport-limited basins, where the frequency and magnitude of débris flow events are controlled primarily by hydroclimatic triggering [see also *Carson and Kirkby*, 1972]. In many cases, the interval from the occurrence of the last débris flow can help in estimating the magnitude of the next event [*Jakob et al.*, 2005], if activations of new sediment sources at the basin scale due to extreme events are excluded (e.g., widespread landslides). The recharge of sediments occurs mainly along the débris-flow transit channel, where sediment storage and availability can be inferred by cycles of aggradation and degradation [*Jakob et al.*, 2005; *Fuller and Marden*, 2010; *Berger et al.*, 2011].

7. Sediment-Transport Capacity of Glaciers

The concept of transport capacity has not traditionally been prominent in glacial geomorphology. Indeed, the term has been so poorly adopted by the glaciological community that it has sometimes been used in quite different ways. For example, *Hagen* [1987] uses transport capacity to describe the potential flux of ice through a surging glacier system, focusing on the glacier as a transporter of ice, rather than as a transporter of sediment.

Nevertheless, *Boulton* [1975], arguing that the lack of a theoretical framework for sediment transport by glaciers was inhibiting progress in the field, sets about providing such a framework for temperate glaciers. Arguing from *Glen* [1952] and *Weertman* [1957, 1964], he derived an expression for the "transporting power" of a glacier as

$$P_t = \frac{V_{p'} V_i}{(V_i - V_{p'})} \frac{\left[\tau - \left(\frac{V_i}{A R^4} \right)^{0.5} \right]}{N \mu'} \tag{37}$$

where $V_{p'}$ = is the mean velocity of the débris entrained by the glacier (m a^{-1});
N = effective pressure (N);
V_i = basal ice velocity (m s^{-1});
R = bed roughness (−);
μ' = average coefficient of friction (−).

The coefficient A is given as

$$A = \left(\frac{KCB}{L\rho_i}\right)^{\frac{1}{2}}$$

(38)

where L = latent heat of fusion of ice ($J\,g^{-1}$);
K = thermal conductivity of ice ($W\,m^{-1}\,K^{-1}$);
ρ_i = ice density ($kg\,m^{-3}$);
B = the constant in Glen's flow law ($-$);
C = a constant relating pressure and resultant lowering of the freezing point (-).

However, this kind of work has more often focussed purely on entrainment (or deposition) mechanisms and has rarely given explicit consideration to sediment transport capacity. Iverson [2000] calculated the rate of infiltration regelation into subglacial sediments. For débris-rich ice formed by regelation around obstacles to flow, Lliboutry [1993] calculated the likely thickness of a steady-state layer of regelation ice. Calculations of this type could be put to the task of estimating some aspect of transport capacity, but this is not a direction that has been taken to any significant degree. Subsequent work on sediment-transport mechanisms by ice [e.g., Melanson et al., 2013] is not founded on concepts of transport capacity or power.

One review that does explicitly consider sediment-transport capacities of the glacial system is provided by Alley et al. [1997]. Their focus was on how glaciers entrain and transport basal sediment, and they focus exclusively on the transport capacities of various components of the basal environment: subglacial streams, débris-rich basal ice, and the mobile subglacial débris layer. Although it does provide an illustration of some of the issues facing studies of glacier transport capacity, this approach excludes major areas of the whole-glacier sediment transport system such as the supraglacial and englacial débris transport, which can be very important in many glacier settings. It also raises the problem of defining what we mean by glacier sediment transport: is the deforming subglacial layer part of the glacier? Is a meltwater stream within the ice part of the glacier? What about rock glaciers?

In the analysis by Alley et al. [1997] the concept of an actual capacity is focussed mainly on the water sections of the system. In other parts of the basal system (basal ice and subglacial débris) the focus is on entrainment capacity or potential entrainment rates. Sediment transport by glacial water streams involves essentially the same issues as fluvial sediment transport, except that subglacial environments have special conditions including sections under pressure and highly variable discharge (Figure 11). Alley et al. define transport capacity as "volume of sediment transported per time by the system, for a given distribution of grain sizes if the supply of sediment is not limiting" [Alley et al. [1997, p. 1018]. They consider a range of previous studies of bedload transport including the assessment by Gomez and Church [1989] that Bagnold's [1980] equation best predicted bedload-transport rate. Alley et al. [1997] also consider pipe flow, which is relevant for flow in subglacial tunnels. In glacier systems typically steep gradients with high headwater pressures and high but variable discharges, including high-frequency discharge pulses, make glacial meltwater streams particularly effective transporting agents. Subglacial water flow and sediment processes are very different if surface meltwater can reach the bed, so predicting the capacity of the system would depend on knowing about the connectivity of surface, englacial, and basal drainage networks. This connectivity is likely to be variable over time and will evolve on a seasonal basis. It will also vary between glaciers. Nevertheless, Alley et al. [1997] do review a substantial set of potentially useful approaches to identifying at least some notion of a transport capacity for the fluvial components of the glacial system. However, because glaciers involve four-phase flow (water, ice, air, and sediment), it is unlikely that any simple approach to estimating transport rate or capacity could ever be achieved.

For other components of the basal sediment system, however, Alley et al. [1997], and others before and since, have greater difficulty even approaching a theoretical prediction of sediment fluxes. For example, on the question of predicting sediment transport through the deforming subglacial layer [Alley et al. [1997, p. 1021] suggest that "without an accurate flow law including controls on viscosity, one cannot calculate débris fluxes from first principles." On predicting sediment transport by glaciotectonic mechanisms they write "Despite the fact the glaciotectonic structures are common in glaciated regions... no general theory has been advanced to account for the amount of debris transported in front of or beneath glaciers in this way." As for transport of débris within basal ice, there is still no clear agreement even on the processes by
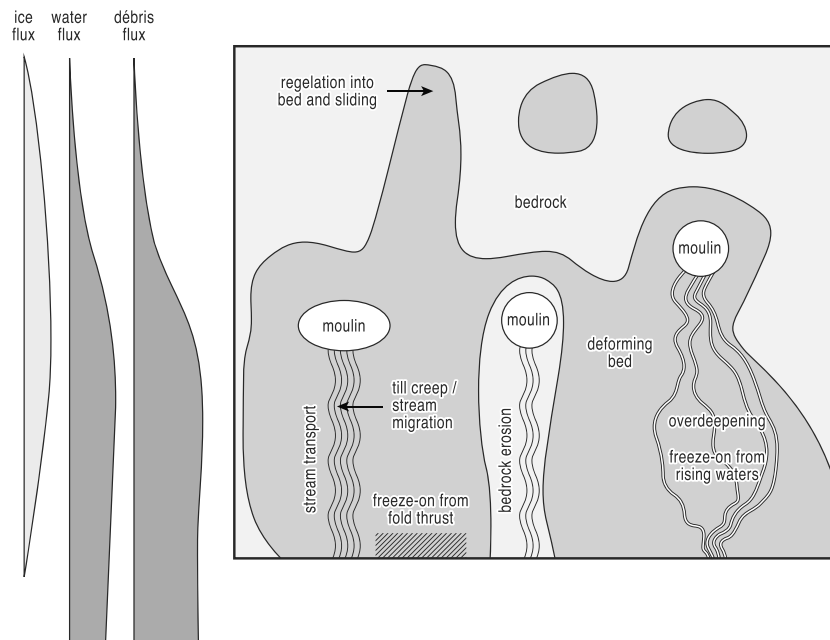
Figure 11. Schematic diagram of possible subglacial conditions, showing a plan view of the bed of an active ice sheet or glacier flowing from the top to the bottom of the page. The key sediment transport mechanisms are illustrated. The thicknesses of the curves on the left represent the ice flux (linear scale), the water flux (logarithmic scale), and total flux of débris in the ice plus deforming or stream-transported sediment below the ice (logarithmic scale) [after *Alley et al.*, 1997].

which débris can be entrained, let alone any quantitative estimation of the potential débris flux through the basal layer. Several studies [e.g., *Hunter et al.*, 1996; *Knight et al.*, 2002] have measured actual débris flux through the basal layer by combining measurements of débris content and ice velocity, but these studies have not extended to identifying theoretical limits. The aim of these studies has been to explore how much débris glaciers *do* carry, not how much they *can* carry.

One reason for ignoring the concept of capacity might be that glaciers are immensely complicated systems for which we still lack an understanding of many fundamental processes. There is still neither a clear flow law for glaciers [*Alley*, 1992] nor a comprehensive understanding of the mechanisms of glacial débris entrainment and deposition [*Evenson and Clinch*, 1987; *Alley et al.*, 1997]. Both would be required for an analysis of glacier transport capacity, which would need to address many variables beyond a simple fluid mechanics of ice-débris interaction. *Evenson and Clinch* [1987, p. 112] suggested that "glacier margins are probably the single most complicated clastic sedimentation system and that there are [sic] a bewildering array of potential sediment transport paths within any glacier system". *Alley et al.* [1997, p. 1030] argued that "the complexity of glacial systems almost entirely precludes succinct rules describing erosion and sedimentation." Sediment transport in glaciers is complicated because there are so many different transport paths within the glacier system (supraglacial, englacial, basal, subglacial, and fluvioglacial), so many débris sources, so many mechanisms of débris transport, such a huge range of particle sizes, and so many processes by which débris can be entrained and released, few of which relate to the fluid-flow properties of the ice. In many of these areas there is still no agreement even about the fundamental operation of processes. For example, the significance of subglacial glaciohydraulic supercooling as a mechanism for entraining substantial amounts of sediment is still being evaluated, and the details of the environmental factors that control the mechanism are still being explored [*Alley et al.*, 1998; *Cook et al.*, 2009; *Creyts et al.*, 2013]. Therefore, it would not yet be straightforward to incorporate this process, which may be extremely important, into any working model of entrainment potential, let alone one of transport capacity. We do not yet have all the parts to put together into a comprehensive model of glacier sediment transport. Arguably, transport capacity is not prominent in glacial geomorphology because the science is not yet sufficiently advanced to address the problem.

The questions that arise in glaciology and glacial geomorphology surrounding débris transport do not center around the capacity of the system. There are many other issues being considered, and sediment plays an

important role in glacier dynamics, but transport capacity is not often close to the forefront of thinking within the discipline. Nevertheless, glacial studies have considered factors similar to those considered by geomorphologists looking at fluvial and aeolian transport capacity. Ice velocity, basal shear stress, effective normal pressure, and the susceptibility of basal material to the tractive force applied by the movement of the overlying material are central to many aspects of glacier sediment dynamics and have been treated in detail by glaciologists and glacial sedimentologists. In particular, studies of how subglacial sediments may be mobilized by stress imparted from the ice above come closest to linking ice flow with sediment transport in a way similar to fluvial or aeolian studies.

8. Discussion

Transport capacity as a concept originally developed in fluvial systems and has received the greatest and most enduring attention in this process domain. Gilbert introduced the term in his study that aimed to address "the common needs of physiographic geology and hydraulic engineering" [*Gilbert*, 1914, p. 9] and was directly influenced by engineers from France and Germany. The need to underpin engineering decisions runs across a range of aspects of the development of the concept in fluvial, aeolian, coastal, and hillslope geomorphology and is reflected by common practice in all of these areas (Table 1), even if uncertainty in results (see Figure 1) more often than not leads to calibration and uncertainty of how the concept does actually justify any of the results obtained. The idea that there should be a single value of sediment transport that corresponds to a particular discharge also underpins the regime theory developed by engineers to design stable channels over a similar time period to the original work on transport capacity [e.g., *Kennedy*, 1895; *Lacey*, 1930]. However, it is intriguing to note that there seems not to have been a direct link between these literatures, at least until much later when the work on regime theory started to inform broader concerns of river management and restoration [*White et al.*, 1982; *Hey and Thorne*, 1986]. This parallel development is also reflected in the apparent parallel inventions in the different subfields of geomorphology, although authors were often quick to point out the linkages (e.g., *Einstein* [1941] and his discussants [*Einstein*, 1942] recognized the link between the work on fluvial bedload and the work of *Bagnold* [1936] in the aeolian realm, even if the recognition was not mutual until much later). Although many of these developments have been related to positivist and deterministic approaches relating to prediction, there is also an ongoing recognition that there is a need to engage with aspects of variability. For example, *Einstein* [1941, p. 561] noted the need to base these predictions on the "new theories of turbulence" of *Shields* [1936] and *Rouse* [1939]. However, this need to address ongoing developments in the field of turbulence has not always been heeded (see discussion below). It also fails to recognize the conceptual limitations that may mean that there is no single concept of transport capacity that is applicable across all areas of the discipline, nor indeed to individual process domains under all conditions. Consequently, there are issues both in terms of whether any predictions based on the concept have a real, mechanistic understanding and also in terms of how concepts of sediment transport are communicated between different parts of the discipline or to those working in related areas in interdisciplinary projects.

Although the development of these different process domains can be considered as developments within geomorphology, they were principally carried out by engineers (Einstein was a hydraulic engineer; Meyer and Wischmeier were agronomic engineers; Caldwell worked for the Beach Erosion Board (now morphed into the Coastal Engineering Research Centre run by the U.S. Army Corps of Engineers); Gilbert, although having a geological background, saw his research as having engineering applications; and Bagnold was something of an anomaly, working at the time as an independent researcher [*Bagnold*, 1990], although his results were quickly put to practical application). The shift to the implementation of these engineering-based approaches in geomorphology came with the development of quantitative geomorphology in the 1950s and broader notions of the need for quantification and prediction. One of the main proponents of this development, Strahler, was clearly aware of Gilbert's definition of transport capacity and the work on which it was based "Had Gilbert's philosophy of physical geology prevailed among students of landforms the analysis of slopes would not have been so long delayed. … An inevitable result … has been a gradual reduction in the physical science background of geomorphology students and teachers; and a consequent general sterility in original geomorphic research. But while academic geomorphology has been approaching stagnation important developments in the understanding of slope erosion processes have been made by hydrologists, hydraulic engineers, and soil erosion specialists concentrating upon soil conservation and sedimentation engineering" [*Strahler*, 1950, p. 210].

Table 1. A List of Commonly Used Models That Employ Sediment-Transport Capacity Relations Across Different Process Domains

Model	Transport-Capacity Equations	Reference
HEC-RAS (Hydrologic Engineering Center's River Analysis System)	Fluvial erosion: *Ackers and White* [1973] *Engelund and Hansen* [1967] *Laursen* [1958] *Meyer-Peter and Müller* [1948] *Toffaleti* [1968] *Yang* [1973, 1984] *Wilcock* [2001]	*Brunner* [2010]
ISIS: river and floodplain modeling	Fluvial transport: *Ackers and White* [1973]	*CH2MHILL* [2015]
DELFT3D: 3-D modeling suite to investigate hydrodynamics, sediment transport and morphology, and water quality for fluvial, estuarine, and coastal environments	Wave erosion: *Bijker* [1967, 1971] *Soulsby* [1997] *van Rijn* [1993] Current erosion: *Ashida and Michiue* [1972] *Engelund and Hansen* [1967] *Meyer-Peter and Müller* [1948] *Wilcock and Crowe* [2003]	*Deltares* [2014]
MIKE21: simulation of physical, chemical, and biological processes in coastal and marine environments	Current erosion: *van Rijn* [1993] *Engelund and Fredsøe* [1976] *Engelund and Hansen* [1967] *Meyer-Peter and Müller* [1948]	*DHI* [2013]
WEPP (Water Erosion Prediction Project)	Hillslope erosion: *Foster* [1982] based on *Yalin* [1963]	*USDA* [1995]
TOPMODEL (Topography based hydrological MODEL)	Hillslope erosion: *Kirkby* [1993]	*Kirkby* [1997]
EUROSEM (European Soil Erosion Model)	Splash erosion: *Poesen* [1985] *Govers* [1991] *Everaert* [1992] *Poesen and Torri* [1988] Rill erosion: *Govers* [1990] Interrill erosion: *Everaert* [1991] Fluvial erosion: *Govers* [1990]	*Morgan et al.* [1998]
KINEROS (Kinematic Runoff and Erosion Model)	Hillslope and channel erosion: *Ackers and White* [1973] *Engelund and Hansen* [1967] *Kilinc and Richardson* [1973] *Meyer and Wischmeier* [1969] *Yalin* [1963] *Yang* [1973]	*Woolhiser et al.* [1990]

Once the idea made it into the discipline, further developments of the concept within the fluvial and (to a lesser extent) aeolian contexts went on to inform approaches within the coastal domain, soil erosion, and in débris flows and glaciers to a much more limited extent. As the transfer from one area to another occurred, there seems to have been restricted questioning of the core concept, and in many settings (especially in fluvial and slope studies) it seems to have become "black boxed" [*Latour*, 1987] and immune from critical evaluation. For example, *Wainwright et al.* [2008] discuss how the soil-erosion literature has avoided a critical evaluation of not only transport capacity but also other issues such as the mode of sediment transport on hillslopes.

That there seem to have been multiple, independent inventions of the concept is not unusual (e.g., as with evolution, calculus, and the periodic table). Indeed, there is a close parallel in multiple inventions of the notion of carrying capacity by game and range managers, population biologists, ecologists, and

demographers [*Sayre*, 2008]. All have their origins in the broader concept of engineering capacity (and what came to be known as payload of ships) and relate to idealized, static conditions. *Sayre* [2008] pointed out that all developed to relate to the control of environmental systems but have all become difficult to sustain when issues of scale, variability, and system dynamics were addressed. These aspects will be considered further in relation to transport capacity below, not least because of the parallel in the idea of control based on the engineering underpinnings of the term.

In fluvial geomorphology, transport capacity is still very central despite the critiques discussed here in relation to turbulent processes. The importance of the concept has largely remained because of two reasons. First, it is has proven useful in river restoration, channelization, and management for estimating the channel geometry required to transport sediment in design flows [e.g., *White et al.*, 1982]. If the capacity estimate gives an overestimate of the maximum transport rate, although the engineering of structures in the channel will be more expensive, the structures would be secure beyond the design criterion. Secondly, the transport equations that have evolved due to the concept utilize river flow and channel variables that are commonly monitored. However, what is easily measured is not necessarily what is required in terms of the parameters that actually control sediment transport resulting in what are effectively empirical equations not rooted in physical understanding. The usefulness of the approach is thus more about being able to justify practical applications with recourse to an extended heritage of literature, rather than with a clear demonstration of understanding of the process, as would be required by a realist approach.

In the case of aeolian geomorphology, the decline of the concept of transport capacity may, in part, be related to the typical condition in the field of supply limitation. It was recognized early on in field work that the sediment transport in locations where the source material was not just composed of sand was not "at capacity" or "saturated" because so often very little sediment was observed to be moving [*Gillette et al.*, 1980, 1997; *Gillette and Chen*, 2001]. As information about turbulence filtered into the area from fluvial research [e.g., *Bennett and Best*, 1995, 1996; *Jackson*, 1976], aspects of turbulence controls on observed transport meant that spatial and temporal variability suggested that there was not a single capacity value that had a meaning [e.g., *Baas and Sherman*, 2005].

In coastal studies the concept of transport capacity has been implicitly subsumed into the notion of longshore transport rate, which is still an important element in understanding beach development. It is applied more at an annual to decadal timescale whereby the general field state of a beach system might be gauged in terms of sediment input and transport potential out. However, the research emphasis has now switched to the self-regulation of beach form by tuning incoming incident waves and secondary currents post–wave breaking to short-term movement of sediment into distinctive beach morphologies where both cross-beach and longshore beach transport may be codominant or negligible. This emphasis gave rise to morphodynamics [*Short*, 1999b] as the principal mode of beach analysis over the last two to three decades. Transport equations are still considered, but analysis is now more dependent on the detailed spectral breakdown of processes correlated with detailed empirical sediment transport from acoustic sampling (sand grades). This means that detail of sediment transport as a result of power potential is often sidestepped. *Green and Coco* [2014] have argued that in the coastal process domain (as well as argued above in the aeolian), the reason is commonly to do with supply limitation.

In hillslope studies, the concept is only just being reevaluated. *Wainwright et al.* [2008] pointed out the limitations of extrapolation of techniques from the fluvial literature to hillslopes that were untested because of the difficulties of measuring the rapidly changing conditions in very shallow flows and because a number of the fundamental assumptions are not met in those flows. There has been some debate about these arguments [e.g., *Smith et al.*, 2010; *Wainwright et al.*, 2010], but the community seems largely happier to calibrate existing models than to address the fundamental basis of why that calibration is necessary (as indeed is the case in fluvial examples [e.g., *Xia et al.*, 2013]). The use of the concept is also starting to fall out of favor with coastal geomorphologists because of the amount of calibration required to produce predictions at any location. As noted above, apart from some fluvially inspired dabbling, neither glacial nor debris flow geomorphologists have been much influenced by the concept, not least because of issues of complexity and variability.

The foregoing review demonstrates that much of the literature rests on the assumption that a specific, unchanging capacity to transport sediment exists. To what extent is such an assumption reasonable? Where the concept has been developed (and redeveloped) across the different areas of geomorphology,

there has been a convergence to the idea that the *power* of the transporting medium uniquely allows the prediction of the amount of sediment that will be moved. Presumably, the idea of a single power term was developed to avoid complications from other sources of flow variability, but as has been seen, this simplification has proved problematic. If that conceptualization held, then it would provide a powerful tool in the management of environmental systems or in the prediction of landscape evolution. Notwithstanding the lack of consensus on how that power should be estimated, there are a number of conceptual issues that call into question whether such a simple definition exists, and therefore if there are multiple definitions, whether the core concept remains testable as opposed to the evaluation of ancillary hypotheses that simply protect that core from critique. These conceptual issues are as follows: (i) whether or not the power of a flow can be characterized independently of the sediment transported by that flow and whether the power represents adequately the ways in which turbulent structures in different process domains control sediment transport, (ii) the complicated nature of transport systems in any geomorphic domain, and (iii) whether or not there is a scale independence of measured rates of transport.

Perhaps most importantly, the nature of flow is not independent of the sediment transported by that flow. As the flow starts to take on more of a two-phase nature (or even three phase when air is rapidly trapped in water-sediment mixes, for example, in flood waves in ephemeral channels, on beaches or in débris flows, or four-phase flows when ice is included in glacial systems), the density and viscosity change so that the underlying assumptions of the equations used to make predictions diverge from the conditions in the field. In the fluvial and hillslope domains, these changes ultimately lead to the production of hyperconcentrated and then ultimately débris flows, which demonstrate non-Newtonian behavior. Although this point is also recognized by *Hessel and Jetten* [2007], they overlook the practical implication that it means that it is conceptually impossible to parameterize a model based on a single capacity value. In aeolian-dominated systems, it has long been recognized that the initial movement of sediment by the air on its own requires higher rates of wind shear than initial movement when the air contains sand grains that impact the surface during rebound in saltation (called the fluid and impact thresholds, respectively) [*Bagnold*, 1941, pp. 92–95]. The Owen effect also describes how the saltating sediment affects the effective roughness length of the surface, thus changing the transport behavior of the system [*Owen*, 1964]. It is usually argued that this process is insignificant in the fluvial domain because of the lower difference between the density of the sediment and the fluid [*Bagnold*, 1973, p. 484] and is of course irrelevant by the time the flow has become non-Newtonian. *Long et al.* [2014] have shown that in rainsplash there can be an important effect of ballistic impact mobilizing grains as initially splashed particles impact on the surface. Only as techniques develop to look at the dynamics of multiphase flow in more detail will it be possible to evaluate to what extent these secondary effects of particle-initiated entrainment are more than noise.

A further issue is the complicatedness of transporting systems. As already noted, this complicatedness means that the concept of transport capacity has provided limited explanatory power in glacial geomorphology or indeed on hillslopes. The variability in processes in other domains may also be an ultimate reason for the limitation. For example, on hillslopes, the variation between raindrop detachment and splash, flow detachment and flow transport introduces a wide range of spatial heterogeneity in observed transport [*Parsons et al.*, 2004, 2006] so that spatial heterogeneity of surface and subsurface properties will introduce significant variability in transport. Even if it were possible to make a single prediction of capacity, it might be so difficult to account for the underlying stochastic nature of surface properties to render the practical application useless (e.g., because the range of potential values is extremely broad, as suggested by the envelope curves in Figure 4) or as discussed above that it would produce a significant overestimate leading to increased engineering costs. Furthermore, the nature of transport also varies from creeping to rolling to saltation to suspension, which will affect both the ability of the flow to transfer energy to particles and the effect of particle-particle collisions in affecting the energy of the flow. The thresholds between these different forms of transport are not necessarily clear [*Parsons et al.*, 2015], further calling into question the predictability of the system. Relations derived in one domain of transport cease to hold in others in part because of the variability in mechanisms and the difficulty of evaluating which mechanisms are in operation at which points in time and space. In coastal studies, the complexity occurs as a result of multidirectional flows whether offshore under oscillating waves or onshore where swash and backwash are at odds with each other. Consequently, transport power is often used as an explanatory variable in beach transport (i.e., relating to the potential power of a breaking wave), but the transformation postbreaking on the beach face becomes a more complex

issue that means for most geomorphologists, it is easier to talk about macroscale or bulk changes rather than the reductionist statements about instantaneous power. Indeed, it is often stated that it is the hiatuses in flow that cause the main interpretation problem for modeling long-term coastal response. Again, it is not necessarily the most straightforwardly measured variables that are the most significant in explaining patterns of sediment transport.

If based on the assumption that there is a specific, unchanging capacity to transport sediment, there is a further fundamental problem with the concept of transport capacity, in that in order for it to hold, a flow must exhibit the same transport capacity at different temporal and spatial scales. In other words, transport capacity must be independent of the scale of measurement, or there needs to be an explicit representation of spatial and temporal variability in the prediction, above and beyond the flow power terms. Evidence from studies of sediment transport on hillslopes, within rivers, by wind and along coasts suggests otherwise. This evidence can broadly be considered in terms of the temporal and the spatial variability of transport.

Sediment transport is a stochastic process, whereas the transport capacity concept is essentially deterministic. Especially for low concentrations, sediment transport is a granular phenomenon [*Cooper et al.*, 2012] so understanding it has to be at the grain scale [see also *Furbish et al.*, 2012]. Because of stochasticity, transport capacity varies with temporal scale and can only reflect certain time-averaged (steady state) conditions. The evidence for the dependency of transport rate on temporal scale is as follows.

First, fluctuations in sediment transport under quasi-steady flow conditions occur commonly in rivers, hillslopes, and aeolian transport. These sediment pulses occur over a range of temporal scales: (i) turbulence timescales due to the advection and propagation of turbulent-flow structures (e.g., *Drake et al.* [1988] and *Radice et al.* [2013] for rivers and *Baas and Sherman* [2005] and *Durán et al.* [2011] for aeolian transport); (ii) timescales of tens to hundreds of seconds due to wind gusting [*Baas*, 2004; *Butterfield*, 1998], variability in wave power, migration of bed forms (e.g., *Cudden and Hoey* [2003], *Iseya and Ikeda* [1987], and *Whiting et al.* [1988] for rivers, *Andreotti et al.* [2002a, 2002b, 2010] and *Sauermann et al.* [2001] for aeolian transport, and *Kosov et al.* [1978] (cited in *Sidorchuk* [1999]) for hillslopes), or changes in the size and structure of river bed material [e.g., *Gomez*, 1983; *Pender et al.*, 2001]; and (iii) at hourly scales in rivers due to processes that scale with the width of the channel, such as bar migration [*Gomez et al.*, 1989], bank erosion [*Cudden and Hoey*, 2003], scour-fill sequences, and changes in sediment supply [e.g., *Gilbert*, 1917; *Knighton*, 1989]. Similarly, coastal sediment transport is now seen as being dominated by the concepts of secondary generation of periodic longshore and onshore currents generated by variable amplification of the shoaling incident wave spectrum [*Huntley et al.*, 1977; *Short*, 1999b; *Aagaard et al.*, 2013]. Infragravity wave energy—i.e., those of a lower frequency than gravity waves—as well as incident wave energy is variably experienced on a beach face, depending on beach face reflective-dissipative status (often now indexed by the surf-similarity parameter [*Battjes*, 1974], which is a function of the overall incident wave steepness (wave height over wavelength) relative to overall shoaling shoreface slope [*Huntley et al.*, 1977; *Wright et al.*, 1979; *Hughes et al.*, 2014]). Consequently, transport equations of the form proposed by Caldwell are only ever broad time averaged at best, as well as being poor indicators of overall transport variability.

Secondly, these fluctuations in sediment transport imply that estimates of transport capacity are dependent upon sampling duration. For example, if particle movement is observed over a longer time frame, one would expect a higher likelihood of measuring large transport distances because more particles are transported. A particle that has traveled further may well be deposited in a more stable position than those that have traveled a shorter distance. For example, in rivers, *Ferguson et al.* [2002] have demonstrated an apparent deceleration of tagged movement of fluvial gravel through a succession of floods, which may relate to the structure of bed material. *Wainwright and Thornes* [1991] and *Parsons et al.* [1993, 2010] make similar observations for particles moving on hillslopes following multiple storm events, as do *Kirkby and Statham* [1975] and *Statham* [1976] for rockfalls. Given that sediment flux is a product of the entrainment rate and transport distance, it follows that the sampling duration affects estimates of sediment flux and thus estimates of the transport capacity of the flow [*Bunte and Abt*, 2005; *Furbish et al.*, 2012]. In addition, sediment pulses that occur as the movement of individual grains exposes others to the flow [e.g., *Cudden and Hoey*, 2003; *Drake et al.*, 1988] are more likely to occur over longer sampling durations.

Thirdly, estimates of transport rate are dependent upon the frequency at which transport is sampled. If one were to observe the movement of transported particles at a temporal frequency that was comparable to the

frequency of turbulent-flow structures, single-entrainment events would be observed. Thus, entrainment rate and transport distance, and therefore their product, which gives sediment flux [*Parsons et al.*, 2004], would be estimated directly. If particle motion were sampled infrequently (e.g., before and after a single flow, storm, or wind event), it would not be possible to determine whether the measured movement is due to single or multiple transport events. Thus, the entrainment rate could not be measured, and the sediment flux would be estimated indirectly by the virtual velocity (ratio of distance traveled to sampling duration), as, for example, in equation (1). Thus, the two estimates would likely give different values for the transport capacity of the flow.

Fourthly, sediment transport is often assumed to adapt to local conditions instantaneously, and accordingly, sediment-transport rate is modeled as a capacity value based on some local flow condition. The fluvial literature has clearly demonstrated phase differences between flow and sediment-transport rate for both suspended and bedload transport [e.g., *Alexandrov et al.*, 2003; *Kuhnle*, 1992; *Laronne and Reid*, 1993; *Lee et al.*, 2004; *Marcus*, 1989; *Reid et al.*, 1985; *Sutter et al.*, 2001; *Williams*, 1989]. The deviation between bedload-transport rate and capacity increases with time lag, and the time lag is longer for higher transport rates [*Cao et al.*, 2010]. As highlighted earlier, a similar situation also exists with wind-blown transport [*Butterfield*, 1991; *Hardisty*, 1993; *Spies et al.*, 2000], in which the time lag is greater with an increase in wind speed than a decrease. Thus, suspended and bedload transport in fluvial and aeolian environments cannot adapt to local flow conditions faster than changes in flow due to turbulence [*Cao et al.*, 2007, 2010; *Stout and Zobeck*, 1997], and hence, the transport-capacity concept cannot provide a mechanistic understanding of sediment transport at turbulence timescales, nor at scales smaller than the time lag.

Fifthly, transport rates change with the duration of the flow event. As a surface erodes, a grain is likely to come to rest in a more sheltered position than its original location so a higher shear stress will be required to remobilize the grain, increasing its subsequent resting duration and travel distance, and virtual velocity. In rivers, transport rates fall to negligible levels, even during steady flows, due to bed armoring [e.g., *Pender et al.*, 2001]. *Hairsine and Rose* [1992a, 1992b] suggest that the probability of detachment on hillslopes is dependent upon whether the particle resides in the deposited layer or in the unshielded original soil layer. Furthermore, in rivers, flow history also influences transport rate. Both subthreshold flows [e.g., *Paphitis and Collins*, 2005] and above-threshold flows [*Hassan et al.*, 2006; *Mao*, 2012; *Reid et al.*, 1985] have been shown to increase detachment thresholds for sands and gravels over a succession of flow events.

In terms of spatial dependency, transport rates are dependent upon the spatial distribution of fluid shear stress and critical shear stress. Thus, transport capacity is dependent spatially. This dependency can be demonstrated in the following ways.

First, spatial variations in transport rates occur at a range of scales; from the grain scale due to differences in the surface microtopography, such as packing and grain exposure [e.g., *Darboux and Huang*, 2005; *Drake et al.*, 1988; *Jackson et al.*, 2006; *Radice et al.*, 2009; *Wainwright and Thornes*, 1991], at the bed form scale [e.g., *Baas and Sherman*, 2005; *Brayshaw*, 1984; *Church et al.*, 1998; *Farres*, 1987; *Gares et al.*, 1996; *Laronne and Carson*, 1976; *Richards and Clifford*, 1991; *Torri*, 1987], and at the plot [e.g., *Vandaele and Poesen*, 1995; *Favis-Mortlock et al.*, 2000] or reach scale [e.g., *Hooke*, 1975; *Dietrich and Smith*, 1984] due to changes in morphology [*Bauer et al.*, 1996; *Gillies et al.*, 2006; *Gillies and Lancaster*, 2013]. Thus, estimates of transport capacity vary according to the spatial scale over which the measurements of transport rate are made. This dependency on sampling area can also be nicely illustrated with the following evidence. Measurements of travel distances of individual particles during runoff events on hillslopes and of tagged gravel in rivers show that transport distances are small and have a heavy-tailed distribution [*Hassan et al.*, 1991; *Wainwright and Thornes*, 1991; *Parsons et al.*, 1993; *Hill et al.*, 2010; *Lajeunesse et al.*, 2010]. Thus, only the smallest eroded particles, or a fraction of larger ones, are likely to be transported to large distances, after even a very large runoff or flow event. Hence, estimates of sediment flux, and therefore transport capacity, will vary with sampling area. The spatial dependency between roughness scale and aeolian transport can also result in a given wind condition not producing the same transport rate at different spatial scales. For equivalent roughness (as defined by roughness density), the size of the roughness elements dramatically affects the transport. For the same shear velocity one can get quite different flux rates as the saltating particles interact with the roughness [*Gillies et al.*, 2006; *Gillies and Lancaster*, 2013].

Thus, although the shear velocity may be the same, the flux may be very different due to the size of the roughness, suggesting that in cases where sediment supply is not limiting, there would either have to be two capacity states for the same shear velocity or that the concept of capacity is not a mechanistic description of the process.

Secondly, transport-capacity equations are not scalable from one river to another, nor from one hillslope to another. Transport capacity is scaled by, among other things, bed shear velocity (or bed shear stress) and grain size. Two flows can have the same bed shear stress but occur over rills or channels with differing slopes and therefore have differing relative submergence (flow depth: bed roughness size). To illustrate this difference, consider the following example. Take two river channels with a slope of 0.005 (−) and 0.02 (−), identical bed shear velocity of 0.1 m s^{-1}, and composed of the same material ($D_{50} = 0.02$ m). Given that $u_* = \sqrt{gh_w S}$, the flow depths h_w for the two river channels are 0.20 m and 0.05 m, respectively, so the relative submergence is 10.2 and 2.5. This difference in submergence may be important when extrapolating the transport-capacity relationships from one river to another, or one from hillslope to another, for the following reason. Within the fluvial literature there is strong evidence that the mean fluid shear stress at which sediment is entrained is inversely correlated to relative submergence [e.g., *Bathurst et al.*, 1983, 1987; *Bettess*, 1984; *Shvidchenko and Pender*, 2000; *Mueller et al.*, 2005; *Parker et al.*, 2011] because the structure of the near-bed flow changes with submergence [*Ashida and Bayazit*, 1973; *Graf*, 1991; *Lamb et al.*, 2008; *Cooper*, 2012]. *Wainwright and Thornes* [1991] found a similar pattern for coarse particles on hillslopes. One further problem in trying to use the concept for different slopes is the concept that only applies strictly to steady and uniform flows. Therefore, the degree to which the flow can transport its capacity depends on the steadiness and uniformity of the flow. As slope increases, the flow is more likely to be nonuniform and locally unsteady. Thus, these differences in flow submergence, uniformity, and steadiness on hillslopes or within channels with differing slopes mean that flows with the same mean bed shear stress will not have the same putative transport capacity.

Thirdly, transport capacity equations do not scale across different process domains. If the concept is physically robust, this scaling problem should not arise. For example, the extrapolation of transport-capacity equations from the fluvial literature to processes occurring on a hillslope, though commonplace, poses potential problems. The relative flow submergence under which transport occurs is usually higher for a river so the transport relationships developed for rivers may not scale to overland flow conditions. For a given bed shear velocity and grain size, the slope of a hillslope is likely to be greater than in a river, and the flow depth and relative submergence will therefore be lower. To illustrate this issue, consider the following example. Take a hillslope with a typical slope of 0.2 (−) and a river with a slope of 0.005 (−), an identical bed shear velocity of 0.1 m s^{-1}, and composed of the same material ($D_{50} = 0.0005$ m). The flow depths h_w for the hillslope and river are 0.005 m and 0.2 m, respectively, so the relative submergence is 20 and 815. Therefore, the difference in relative submergence between river and overland flows is likely to result in transport rates on hillslopes being underestimated by fluvial transport-capacity equations. Furthermore, the use of bed shear velocity as the scaling parameter from the river and to the hillslope relies on the assumption of steady and uniform flow. These conditions are unlikely to occur on the higher gradients commonly found on hillslopes. A similar or somewhat analogous situation happens in aeolian transport as affected by roughness scale. Based on results presented in *Brown et al.* [2008] and *Raupach et al.* [2006], for equivalent roughness densities ($\lambda =$ total roughness element frontal area/area occupied by the elements), the same shear stress will be exerted on the surface among the roughness elements, regardless of their size and distribution. The saltation flux for similar roughness densities, however, can be quite different, which *Gillies et al.* [2006] and *Gillies and Lancaster* [2013] attribute to the interaction of the particles in transport with the roughness elements. Large elements reduce the flux beyond that which can be solely attributed to the effect caused by the partitioning of shear stress between the roughness elements and the surface. This observation suggests that there can be multiple apparent capacity states for the same surface shear stress in aeolian sediment transport, due to roughness effects related to their physical size.

If the issues highlighted in this discussion mean that it is difficult to retain a concept of transport capacity, what are the implications for understanding and predicting sediment transport? The experience from especially the fluvial, aeolian, and coastal process domains suggests that a concept of capacity is not a prerequisite for predicting transport rate. If we focus on the grain scale, then employing the terminology of *Furbish et al.*
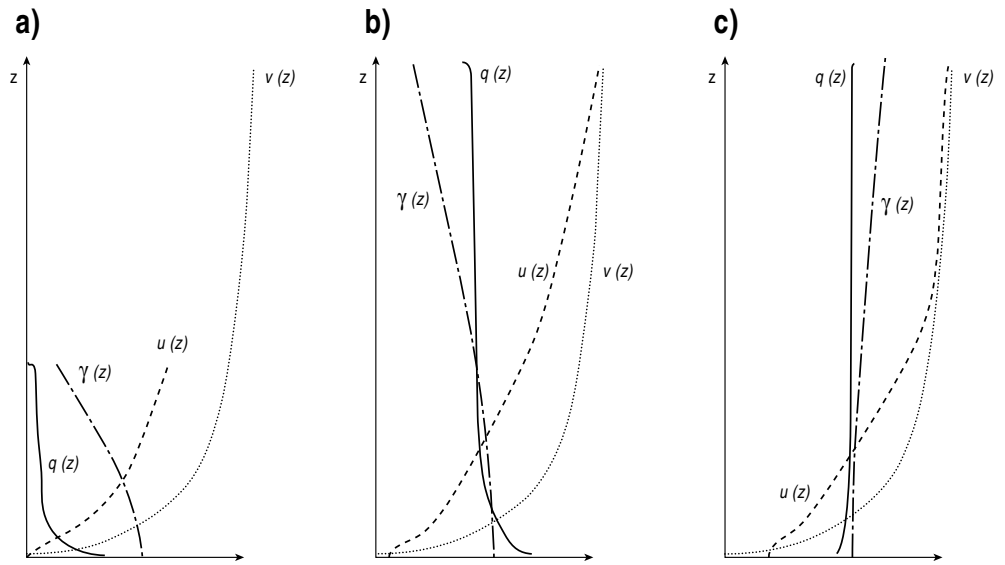
Figure 12. Relationships as a function of flow depth (z) between mean flow velocity (v), mean sediment velocity (u), mean particle activity (γ —see equation (39)), and mean sediment flux (q) for (a) flow dominated by coarse particles moving near the substrate, (b) flow dominated by fine particles moving throughout the flow column, and (c) hyperconcentrated or débris flows where the velocities of the fluid and the sediment converge. In moving from Figure 12a to Figure 12c, the upstream effects of flow variability cause both $u(z)$ and $\gamma(z)$ to become more variable because of anisotropy in the turbulent flow.

[2012], the flux represented by an ensemble of sediment particles at a particular point in space and time can be defined as

$$\overline{q_x}\,(x,\,t)\,=\,\overline{u}\,\overline{\gamma}\,-\,\frac{1}{2}\,\frac{\partial}{\partial x}\,(\overline{\kappa}\,\overline{\gamma}) \tag{39}$$

where $\overline{q_x}$ is the mean unit flux relative to the direction of flow ($L^2 T^{-1}$);

$\bar{u}$ is the mean particle velocity ($L T^{-1}$);

$\bar{\gamma}$ is the mean "particle activity" or volume of particles in the fluid per unit area of the bed (L);

x is distance in the direction of the flow (L);

$\overline{\kappa}$ is the mean diffusivity of particle movement ($L^2 T^{-1}$).

Each of these terms has an implicit scale of variability, and only at low transport rates can they be considered correlated [*Furbish et al.*, 2012]. Particle velocity is controlled by the position of the particle in the vertical profile, which is a function of initial conditions of entrainment and its trajectory, both related to turbulence and to sediment particle size, spin (Magnus force), and interactions with other particles. The particle activity is a function of relative entrainment and deposition rates and thus depends on local conditions in relation to the former but conditions upflow for the latter [see *Parsons et al.*, 2004]. Diffusivity is related to the autocorrelation of variability in the particle velocity [*Furbish et al.*, 2012] and thus again on turbulence structures over distances upflow corresponding to different lengths as a function of different particle sizes. Figure 12 illustrates how these different scales might vary through a vertical flow profile. Transport capacity would only hold if, averaged over a suitable timescale (e.g., over timescales of turbulent variations or of bed form translation [*Furbish et al.*, 2012], the sediment flow into the control volumes at different points in the fluid flow balanced the sediment flow out. However, because turbulence in environmental flows is anisotropic, i.e., the flow structures have a downstream directional preference and thus the mean flow velocity has a gradient (see *Grant and Marusic* [2011] for a useful review), changes in particle activity, velocity, and diffusivity will relate to different space and timescales in the flow field and thus supply into—and the delivery out of— the control volume cannot be steady and thus transport capacity cannot be a meaningful description of the process. Transport capacity could only hold in conditions of homogeneous turbulence (i.e., turbulence has the same structure quantitatively in all parts of the volume, so the velocity fluctuations are random and the mean fluctuation is zero) or isotropic turbulence (the statistical features have no directional
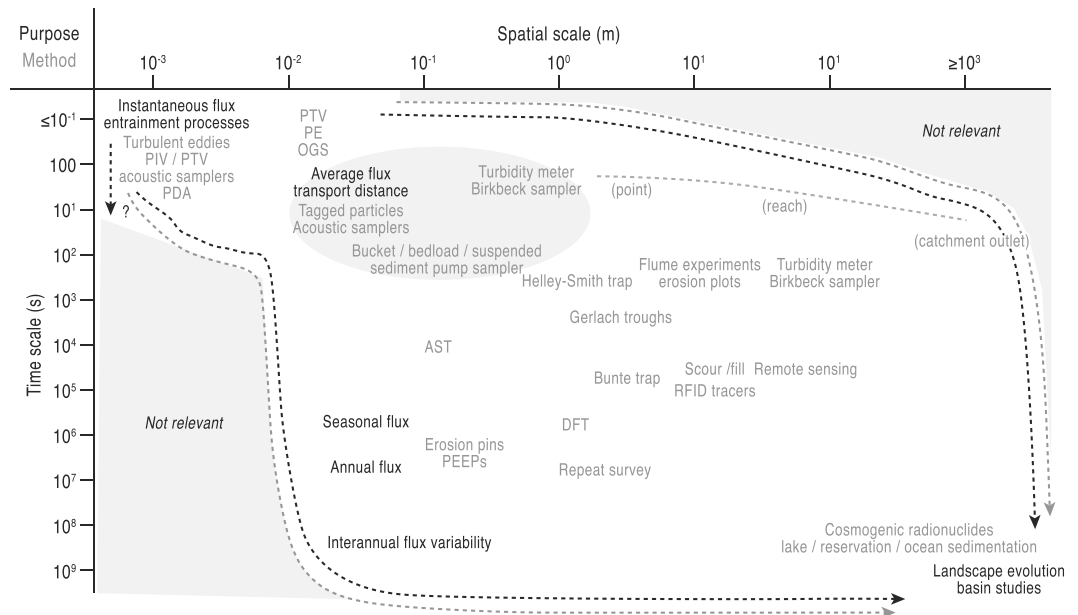
Figure 13. Spatial and temporal scales of measurement of sediment transport and their implicit or explicit uses. The area shaded in the centre of the diagram has been the one typically used to infer transport capacity rates. Measurements at other scales and for other purposes have come to challenge the underlying concept of transport capacity.

preference, and thus, there is no mean flow velocity gradient, and the mean velocity is either zero or constant throughout the volume). As particle activity increases, the probability of particle-particle collisions will increase [e.g., *Bagnold*, 1954; *Sommerfeld*, 2001], the nature of the fluid will change, and thus too will the correlation lengths over which the terms in equation (39) need to be averaged.

The grain-based perspective suggests, then, that what is observed is a time-averaged sediment flux. In this setting, models that relate entrainment to underemployment of capacity or deposition to an exceedance of capacity (e.g., equations (28), (29), (31), (32), and (35)) are physically unreasonable, because they confound the instantaneous processes of transport with a phenomenological representation of time-averaged conditions. Such a perspective also requires a more nuanced approach to the idea of supply limitation. At the very least, there needs to be a recognition of the temporal and spatial variability of sediment supply, which also relates to conditions upflow, in the same way as the variability in transport does. Improved models of sediment transport thus need to focus on direct or indirect characterizations of the terms in equation (39) to allow predictions—whether averaged or including stochastic fluctuations—to be more reliable and transferable because of their process basis. The general nature of equation (39) means that it is a valid basis for developing general, universal models for predicting sediment-transport rates and thus overcome some of the limitations highlighted in this review as a result of working in specific process domains.

This perspective also has implications for what is measured and how those measurements are carried out. At a specific scale, different measurements will reflect different characteristics and states of the sediment-transport process within different components of the geomorphic system. Figure 13 demonstrates how different techniques have been used at different scales. More long-standing techniques tend to emphasize averaging over longer time intervals, and we would argue have tended to support concepts relating to steady state, such as transport capacity, rather than the inherent variability of sediment-transport processes. Newer technologies are increasingly able to evaluate shorter temporal and spatial variability, and it will be will the application of these techniques that the more robust and transferable models of sediment transport will be developed.

9. Conclusions

Transport capacity is a concept that is used across a wide range of domains within geomorphology. Although initially defined in fluvial geomorphology, there were subsequently a number of independent inventions of similar concepts that were subsequently reabsorbed into the discipline. This process happened at an

accelerating pace from the 1950s following the drive to make the discipline more quantitative but was also underpinned from perspectives needed to manage and control environmental systems. The above review suggests that the multiple different ideas and applications of the term add potential for confusion and mean that ways of testing ideas of capacity are unclear. Unless the use in different areas is clarified or common terminology is defined [e.g., *Bracken and Oughton*, 2006], there is a risk that coherent testing of the idea will be impossible. If the fact that the complicatedness of environmental systems and that estimates of capacity do not transfer across scales means that the concept is fundamentally limited, there will be serious practical consequences of using it to make predictions, especially as environmental management increasingly moves toward integrated approaches at catchment or landscape scales (e.g., Integrated River Basin Management and related policies such as the EU Water Framework Directive or aspects of the Clean Water Act in the U.S.). If the existing models need significant calibration and are thus poor representations of the process, the continued use of the model is more about a justification of truth given the need to expend effort to carry out that calibration. For practical use, much simpler models could be calibrated if all that is required is an empirically based prediction of a rate. As noted in the overview on carrying capacity in a range of disciplines by *Sayre* [2008, p. 131] "If carrying capacity is conceived as static, it is theoretically elegant but empirically vacuous; but if it is conceived as variable, it is theoretically incoherent or at best question-begging."

The recognition of the complicatedness of sediment-transport systems and the effects of spatial and temporal dynamics in them—both as a result of turbulence and of environmental heterogeneity—should mean that new approaches are needed that are not underpinned by an ideal transport capacity that is virtually impossible to produce outside of controlled, laboratory conditions. Just because an effect is isolatable in laboratory conditions does not imply that it is a useful way of approaching an understanding of the real world [*Hacking*, 1983]. There is a need to take on board the complicatedness of the environment and of process. In the latter case, there is a strong implication from the comparisons here that an approach that recognizes that different types of flow form continua would be a useful way forward, recognizing that some of the institutional distinctions made in the discipline hinder the development of geomorphological understanding overall. If so, the implication is strongly that we need to move away from the idea of transport capacity and certainly that a single capacity for any set of condition is practically impossible. As the advances discussed above suggest, the way forward requires fundamental characteristics of sediment transport to be reevaluated using an integrated approach that combines both fundamental theory with empirical observations and that the latter should be driven by the former.

References

Aagaard, T., B. Greenwood, and M. G. Hughes (2013), Sediment transport on dissipative, intermediate and reflective beaches, *Earth Sci. Rev.*, *124*, 32–50.

Abbott, J. E., and J. R. D. Francis (1977), Saltation and suspension trajectories of solid grains in a water stream, *Philos. Trans. R. Soc., A*, *284*, 225–254.

Abrahams, A. D., and P. Gao (2006), A bedload transport model for rough turbulent open-channel flows on plane beds, *Earth Surf. Processes Landforms*, *31*, 910–928.

Abrahams, A. D., G. Li, C. Krishnan, and A. J. Atkinson (2001), A sediment transport equation for interrill overland flow on rough surfaces, *Earth Surf. Processes Landforms*, *26*, 1443–1459.

Ackers, P., and W. R. White (1973), Sediment transport: New approach and analysis, *J. Hydraul. Div., Am. Soc. Civ. Eng.*, *99*(HY11), 2041–2060.

Alexandrov, Y., J. B. Laronne, and I. Reid (2003), Suspended sediment concentration and its variation with water discharge in a dryland ephemeral channel, northern Negev, Israel, *J. Arid Environ.*, *53*, 73–84.

Allen, J. R. L. (2001), *Principles of Physical Sedimentology*, The Blackburn Press, Caldwell, N. J.

Alley, R. B. (1992), Flow-law hypotheses for ice-sheet modelling, *J. Glaciol.*, *38*, 245–256.

Alley, R. B., K. M. Cuffey, E. B. Evenson, J. C. Strasser, D. E. Lawson, and G. J. Larson (1997), How glaciers entrain and transport basal sediment: Physical constraints, *Quat. Sci. Rev.*, *16*, 1017–1038.

Alley, R. B., D. E. Lawson, E. B. Evenson, J. C. Strasser, and G. J. Larson (1998), Glaciohydraulic supercooling: A freeze-on mechanism to create stratified, debris-rich basal ice—II. Theory, *J. Glaciol.*, *44*(148), 563–569.

Almedeij, J. H., and P. Diplas (2003), Bedload transport in gravel-bed streams with unimodal sediment, *J. Hydraul. Eng.*, *129*(11), 896–904.

Alonso, C. V., W. H. Neibling, and G. R. Foster (1981), Estimating sediment transport capacity in watershed modeling, *Trans. ASAE*, *24*, 1211–1220 and 1226.

Andreotti, B., and P. Claudin (2007), Comment on "Minimal size of a barchan dune", *Phys. Rev. E*, *76*, 063301.

Andreotti, B., P. Claudin, and S. Douady (2002a), Selection of barchan shapes and velocities. Part 1: Dynamics of sand, wind, and dunes, *Eur. Phys. J. B*, *28*, 321–339.

Andreotti, B., P. Claudin, and S. Douady (2002b), Selection of barchan shapes and velocities. Part 2: A two dimensional modelling, *Eur. Phys. J. B*, *28*, 341–352.

Andreotti, B., P. Claudin, and O. Pouliquen (2010), Measurements of the aeolian sand transport saturation length, *Geomorphology*, *123*, 343–348.

Andrew, E. D. (1979), Scour and fill in a stream channel, East Fork River, western Wyoming, *U.S. Geol. Surv., Prof. Pap., 1117*, Washington, D. C.

Andrews, E. D., and G. Parker (1987), Formation of a coarse surface layer as the response to gravel mobility, in *Sediment Transport in Gravel-Bed Rivers*, edited by C. R. Thorne, J. C. Bathurst, and R. D. Hey, pp. 269–325, John Wiley, Chichester.

Armanini, A., L. Fraccarollo, and M. Larcher (2008), Liquid–granular channel flow dynamics, *Powder Technol.*, *182*, 218–227.

Ashida, K., and M. Bayazit (1973), Initiation of motion and roughness of flows in steep channels, in *Proceedings of the 15th Congress of the International Association for Hydraulic Research*, Istanbul, Turkey.

Ashida, K., and M. Michiue (1972), Study on hydraulic resistance and bedload transport rate in alluvial streams, *Trans. JSCE*, *206*, 59–69.

Ashworth, P. J., and R. I. Ferguson (1989), Size-selective entrainment of bed load in gravel bed streams, *Water Resour. Res.*, *25*, 627–634, doi:10.1029/WR025i004p00627.

Baas, A. (2004), Evaluation of saltation flux impact responders (Safires) for measuring instantaneous aeolian sand transport rates, *Geomorphology*, *59*(1–4), 99–118.

Baas, A. C. W. (2006), Wavelet power spectra of aeolian sand transport by boundary layer turbulence, *Geophys. Res. Lett.*, *33*, L05403, doi:10.1029/2005GL025547.

Baas, A. C. W., and D. J. Sherman (2005), Formation and behavior of aeolian streamers, *J. Geophys. Res.*, *110*, F03011, doi:10.1029/2004JF000270.

Bagnold, R. A. (1936), The movement of desert sand, *Proc. R. Soc. London, Ser. A*, *157*, 594–620.

Bagnold, R. A. (1941), *The Physics of Blown Sand and Desert Dunes*, Chapman and Hall, London.

Bagnold, R. A. (1946), Motion of waves in shallow water: Interaction between waves and sand bottoms, *Proc. R. Soc. London, Ser. A*, *187*, 1–18.

Bagnold, R. A. (1954), Experiments on a gravity-free dispersion of large solid spheres in a Newtonian fluid under shear, *Proc. R. Soc. London, Ser. A*, *225*, 49–63.

Bagnold, R. A. (1956), The flow of cohesionless grains in fluids, *Proc. R. Soc.*, *249*, 235–297.

Bagnold, R. A. (1963), Beach and nearshore processes: Part I. Mechanics of marine sedimentation, in *The Sea: Ideas and Observations on Progress in the Study of the Sea*, vol. 3, edited by M. N. Hill, pp. 507–528, Wiley-Interscience, London.

Bagnold, R. A. (1966), An approach to the sediment transport problem from general physics, *U.S. Geol. Surv. Prof. Pap.*, *422-I*, Washington, D. C.

Bagnold, R. A. (1973), The nature of saltation and of "bed-load" transport in water, *Proc. R. Soc. London, Ser. A*, *332*, 473–504.

Bagnold, R. A. (1980), An empirical correlation of bedload transport rates in flumes and natural rivers, *Proc. R. Soc. London, Ser. A*, *372*, 453–473.

Bagnold, R. A. (1990), *Sand, Wind and War: Memoirs of a Desert Explorer*, 202 pp. Univ. of Arizona Press, Tucson.

Bailard, J. A. (1984), A simplified model for long shore sediment transport, in *ASCE Proceedings of the 19th Conference Coastal Engineering*, pp. 1454–1470, Houston, Tex.

Baldwin, D. S., A. M. Mitchell, and J. M. Olley (2002), Pollutant-sediment interactions: Sorption, reactivity and transport of phosphorus, Agriculture, Hydrology and Water Quality, edited by P. M. Haygarth and S. C. Jarvis, CABI Int., Wallingford, pp. 265–280.

Barry, J. J., J. M. Buffington, and J. G. King (2004), A general power equation for predicting bed load transport rates in gravel bed rivers, *Water Resour. Res.*, *40*, W10401, doi:10.1029/2004WR003190.

Bathurst, J. C. (2005), Effect of coarse surface layer on bedload transport, *J. Hydraul. Eng.*, *133*, 1192–1205.

Bathurst, J. C., W. H. Graf, and H. H. Cao (1983), Initiation of sediment transport in steep channels with coarse bed material, in *Mechanics of Sediment Transport*, edited by B. M. Sumer and A. Muller, pp. 207–213, AA Balkema, Brookfield.

Bathurst, J. C., W. H. Graf, and H. H. Cao (1987), Bed load discharge equations for steep mountain rivers, in *Sediment Transport in Gravel-Bed Rivers*, edited by C. R. Thorne, J. C. Bathurst, and R. D. Hey, pp. 453–477, John Wiley, Chichester.

Battjes, J. A. (1974), Surf similarity, in *Proceedings 14th International Conference on Coastal Engineering*, pp. 466–480, ASCE, Copenhagen, Denmark, doi:10.9753/icce.v14.

Bauer, B. O., D. J. Sherman, and J. F. Wolcott (1992), Sources of uncertainty in shear stress and roughness length derived from velocity profiles, *Prof. Geogr.*, *44*(4), 453–464.

Bauer, B. O., R. G. D. Davidson-Arnott, K. F. Nordstrom, J. Ollerhead, and N. L. Jackson (1996), Indeterminacy in aeolian sediment transport across beaches, *J. Coastal Res.*, *12*, 641–653.

Benett, J. P. (1974), Concepts of mathematical modelling of sediment yield, *Water Resour. Res.*, *10*, 485–492, doi:10.1029/WR010i003p00485.

Bennett, S. J., and J. L. Best (1995), Mean flow and turbulence structure over fixed, two-dimensional dunes: Implications for sediment transport and bedform stability, *Sedimentology*, *42*, 491–513.

Bennett, S. J., and J. L. Best (1996), Mean flow and turbulence structure over fixed ripples and the ripple-dune transition, in *Coherent Flow Structures in Open Channels*, edited by P. J. Ashworth et al., pp. 281–304, John Wiley, Chichester.

Berger, C., B. W. McArdell, and F. Schlunegger (2011), Direct measurement of channel erosion by debris flows, Illgraben, Switzerland, *J. Geophys. Res.*, *116*, F01002, doi:10.1029/2010JF001722.

Berti, M., and A. Simoni (2007), Prediction of debris-flow inundation areas using empirical mobility relationships, *Geomorphology*, *90*, 144–161.

Bettess, R. (1984), Initiation of sediment transport in gravel streams, *Proc. Inst. Civ. Eng.*, *77*, 79–88.

Beuselinck, L., G. Govers, A. Steegen, P. B. Hairsine, and J. Poesen (1999), Evaluation of the simple settling theory for predicting sediment deposition by overland flow, *Earth Surf. Processes Landforms*, *24*, 993–1007.

Beuselinck, L., P. B. Hairsine, G. Govers, and J. Poesen (2002), Evaluating a single-class net deposition equation in overland flow conditions, *Water Resour. Res.*, *38*(7), 1110, doi:10.1029/2001WR000248.

Bijker, E. W. (1967), Some considerations about scales for coastal models with moveable bed, Technical Report 50, WL Delft Hydraulics, Delft, Netherlands.

Bijker, E. W. (1971), Longshore transport computation, *ASCE J. Waterw., Harbors Coastal Eng.*, *97*(ww4), 687–701.

Boise Adjudication Team (2014), Sediment transport in Idaho and Nevada. [Available at http://www.fs.fed.us/rm/boise/AWAE/projects/sediment_transport/sediment_transport_BAT.shtml, last accessed 25th July.]

Boulton, G. S. (1975), Processes and patterns of subglacial sedimentation: A theoretical approach, in *Ice Ages: Ancient and Modern*, edited by A. E. Wright and E. Moseley, pp. 7–42, Seel House Press, Liverpool.

Bovis, M. J., and M. Jakob (1999), The role of debris supply to determine debris flow activity in southwestern B.C, *Earth Surf. Processes Landforms*, *24*, 1039–1054.

Bracken, L. J., and E. A. Oughton (2006), "What do you mean?" The importance of language in developing interdisciplinary research, *Trans. Inst. British Geogr.*, *31*, 371–382.

Bracken, L. J., and J. Wainwright (2006), Geomorphological equilibrium: Myth and metaphor?, *Trans. Inst. British Geogr.*, NS *31*, 167–178.

Brayshaw, A. C. (1984), Characteristics and origin of cluster bedforms in coarse-grained alluvial channels, in *Sedimentology of Gravels and Conglomerates*, edited by E. Koster and R. Steel, pp. 77–85, Can. Soc. Pet. Geol., Ottawa.

Bridge, J. S. (2002), *Rivers and Floodplains: Forms, Processes and Sedimentary Record*, Blackwell Publishers, Oxford.

Brown, S., W. G. Nickling, and J. A. Gillies (2008), A wind tunnel examination of shear stress partitioning for an assortment of surface roughness distributions, *J. Geophys. Res.*, *113*, F02S06, doi:10.1029/2007JF000790.

Brunner, G. W. (2010), *HEC-RAS, River Analysis System Hydraulic Reference Manual*, US Army Corps of Engineers, Davis, Calif.

Buffington, J. M., and D. R. Montgomery (1999a), Effects of hydraulic roughness on surface textures in gravel-bed rivers, *Water Resour. Res.*, *35*, 3507–3521, doi:10.1029/1999WR900138.

Buffington, J. M., and D. R. Montgomery (1999b), Effects of sediment supply on surface textures of gravel-bed rivers, *Water Resour. Res.*, *35*(11), 3523–3531, doi:10.1029/1999WR900232.

Bunte, K., and S. R. Abt (2005), Effect of sampling time on measured gravel bed load transport rates in a coarse-bedded stream, *Water Resour. Res.*, *41*, W11405, doi:10.1029/2004WR003880.

Butterfield, G. R. (1991), Grain transport rates in steady and unsteady turbulent airflows, *Acta Mech. Suppl.*, *1*, 97–122.

Butterfield, G. R. (1998), Transitional behaviour of saltation: Wind tunnel observations of unsteady winds, *J. Arid Environ.*, *39*, 377–394.

Butterfield, G. R. (1999), Near-bed mass flux profiles in aeolian sand transport: High resolution measurements in a wind tunnel, *Earth Surf. Processes Landforms*, *24*, 393–412.

Caldwell, J. M. (1956), *Wave Action and Sand Movement Near Anaheim Bay, California*, *Tech. Mem.*, vol. 68, 21 pp. Beach Erosion Board, US Army Corps of Engineers Washington, D. C.

Cao, Z., Y. Li, and Z. Yue (2007), Multiple time scales of alluvial rivers carrying suspended sediment and their implications for mathematical modelling, *Adv. Water Resour.*, *30*(4), 715–729.

Cao, Z., P. Hu, and G. Pender (2010), Reconciled bedload sediment transport rates in ephemeral and perennial rivers, *Earth Surf. Processes Landforms*, *35*, 1655–1665, doi:10.1002/esp.2005.

Carson, M. A., and M. J. Kirkby (1972), *Hillslope Form and Process*, Cambridge Univ. Press, Cambridge.

Celik, I., and W. Rodi (1991), Suspended sediment-transport capacity for open-channel flow, *J. Hydraul. Eng.*, *117*(2), 191–204.

Cellino, M., and W. H. Graf (1999), Sediment-laden flow in open-channels under noncapacity and capacity conditions, *J. Hydraul. Eng.*, *125*, 455–462.

CH2MHILL (2015), ISIS user manual[Online]. [Available at http://help.floodmodeller.com/isis/ISIS.htm, (Accessed 7th August 2015).]

Charlton, R. (2008), *Fundamentals of Fluvial Geomorphology*, Routledge, London.

Chen, C. (1988), Generalized viscoplastic modeling of debris flow, *J. Hydraul. Eng.*, *114*, 237–258.

Chepil, W. S., and R. A. Milne (1939), Comparative study of soil drifting in the field and in a wind tunnel, *Sci. Agric.*, *19*, 249–257.

Church, M. (2006), Bed material transport and the morphology of alluvial river channels, *Annu. Rev. Earth Planet. Sci.*, *34*, 325–354.

Church, M., and M. A. Hassan (2002), Mobility of bed material in Harris Creek, *Water Resour. Res.*, *38*(11), 1237, doi:10.1029/2001WR000753.

Church, M., M. A. Hassan, and J. F. Wolcott (1998), Stabilizing self-organized structures in gravel-bed stream channels: Field and experimental observations, *Water Resour. Res.*, *34*, 3169–3179, doi:10.1029/98WR00484.

Clayton, J. A., and J. Pitlick (2008), Persistence of the surface texture of a gravel-bed river during a large flood, *Earth Surf. Processes Landforms*, *33*, 661–673.

Cochrane, T. A., and D. C. Flanagan (1996), Detachment in a simulated rill, *Trans. ASAE*, *40*, 111–119.

Cook, S. J., Z. P. Robinson, I. J. Fairchild, P. G. Knight, R. I. Waller, and I. Boomer (2009), Role of glaciohydraulic supercooling in the formation of stratified facies basal ice: Svínafellsjökull and Skaftafellsjökull, southeast Iceland, *Boreas*, *39*, 24–38.

Cooper, J. R. (2012), Does flow variance affect bedload flux when the bed is dominated by grain roughness?, *Geomorphology*, *141*, 160–169.

Cooper, J. R., J. Wainwright, A. J. Parsons, Y. Onda, T. Fukuwara, E. Obana, B. Kitchener, E. J. Long, and G. H. Hargrave (2012), A new approach for simulating the redistribution of soil particles by water erosion: a marker-in-cell model, *J. Geophys. Res.*, *117*, F04027, doi:10.1029/2012JF002499.

Creyts, T. T., G. K. C. Clarke, and M. Church (2013), Evolution of subglacial overdeepenings in response to sediment redistribution and glaciohydraulic supercooling, *J. Geophys. Res. Earth Surf.*, *118*, 423–446, doi:10.1002/jgrf.20033.

Cudden, J. R., and T. B. Hoey (2003), The causes of bedload pulses in a gravel channel: The implications of bedload grain-size distributions, *Earth Surf. Processes Landforms*, *28*, 1411–1428.

Davidson-Arnott, R. G. D., B. O. Bauer, I. J. Walker, P. A. Hesp, J. Ollerhead, and I. Delgado-Fernandez (2009), Instantaneous and mean aeolian sediment transport rates on beaches: An intercomparison of measurement from two sensor types, *J. Coastal Res.*, *S156*, 297–301.

Darboux, F., and C.-H. Huang (2005), Does soil-surface roughness increase or decrease water and particle transfers?, *Soil Sci. Soc. Am. J.*, *69*, 748–756.

Davies, T. R. H. (1988a), Modification of bedload transport capacity in braided rivers, *J. Hydrol. (NZ)*, *27*, 69–72.

Davies, T. R. H. (1988b), *Debris-Flow Surges: A Laboratory Investigation*, *Mitteilungen*, vol. 96, 122 pp., Versuchsanst. für Wasserbau, Hydrologie und Glaziologie, Zurich, Switzerland.

Deltares (2014), *Delft3D-FLOW Simulation of Multi-Dimensional Hydrodynamic Flows and Transport Phenomena, Including Sediments, User Manual*, Deltares, Delft, Netherlands.

De Roo, A. P. J., C. G. Wesseling, and C. J. Ritsema (1996), LISEM: A single-event physically based hydrological and soil erosion model for drainage basins—I. Theory, input and output, *Hydrol. Processes*, *10*, 1107–1117.

Dietrich, W. E., and J. D. Smith (1984), Bedload transport in a river meander, *Water Resour. Res.*, *20*, 1355–1380, doi:10.1029/WR020i010p01355.

Dietrich, W. E., J. W. Kirchner, H. Ikeda, and F. Iseya (1989), Sediment supply and the development of the coarse surface layer in gravel-bedded rivers, *Nature*, *340*(20), 215–217.

DHI (2013), *MIKE21 & MIKE 3 Flow Model FM, Sand Transport Module Short Description*, DHI, Hørsholm, Denmark.

Dong, Z., Z. Zhang, P. Lü, and G. Qian (2011), An aeolian transport model for flat shifting sand fields under dynamic-limiting conditions, *J. Arid Environ.*, *75*, 865–869.

Drake, T. G., R. L. Shreve, W. E. Dietrich, P. J. Whiting, and L. B. Leopold (1988), Bedload transport of fine gravel observed by motion-picture photography, *J. Fluid Mech.*, *192*, 193–217.

Durán, O., P. Claudin, and B. Andreotti (2011), On aeolian transport: Grain scale interactions, dynamical mechanisms and scaling laws, *Aeolian Res.*, *3*, 243–270.

Eaton, B. C., M. Church, and R. G. Millar (2004), Rational regime model of alluvial channel morphology and response, *Earth Surf. Processes Landforms*, *29*, 511–529.

Eaton, B. C., M. Church, and T. R. H. Davies (2006), A conceptual model for meander initiation in bedload-dominated streams, *Earth Surf. Processes Landforms*, *31*, 875–891, doi:10.1002/esp.1297.

Egashira, S., N. Honda, and T. Itoh (2001), Experimental study on the entrainment of bed material into debris flow, *Phys. Chem. Earth Part C*, *26*, 645–650.

Egiazaroff, I. V. (1965), Calculation of nonuniform sediment transportations, *J. Hydraul. Div., Am. Soc. Civ. Eng.*, *91*, 225–247.

Einstein, H. A. (1941), Formulas for the transportation of bed load, *Proc. Am. Soc. Civ. Eng.*, *67*, 351–367.

Einstein, H. A. (1942), Formulas for the transportation of bed load, *Am. Soc. Civ. Eng., Trans.*, *107*, 561–577 (as 1941 paper but with discussion by JW Johnson, AA Kalinske, OG Haywood Jr, S Shulits, JS McNown and HA Einstein, pp. 578–597).

Einstein, H. A. (1950), The bed-load function for sediment transportation in open channel flows, U.S. Dept. of Agric., Soil Conserv. Serv., Tech. Bull., 1026, Washington, D. C.

Einstein, H. A., and N. Chien (1953), *Transport of Sediment Mixtures With Large Ranges of Grain Sizes*, *Missouri River Division Sediment Series*, vol. 2, US Army Corps of Engineers, Missouri River Division, Omaha.

Elbelrhiti, H., P. Claudin, and B. Andreotti (2005), Field evidence for surface-wave-induced instability of sand dunes, *Nature*, *437*, 720–723.

Ellis, J. T., B. Li, E. J. Farrell, and D. J. Sherman (2009), Protocols for characterizing aeolian mass-flux profiles, *Aeolian Res.*, *1*(1–2), 19–26.

Ellis, J. T., D. J. Sherman, E. J. Farrell, and B. Li (2012), Temporal and spatial variability of aeolian sand transport: Implications for field measurements, *Aeolian Res.*, *3*(4), 379–387.

Ellison, W. D. (1947), Soil erosion studies: Part I, *Agric. Eng.*, *28*, 145–146.

Emmett, W. W. (1976), Bedload transport in two, large gravel-bed rivers, Idaho and Washington, in *Proceedings, Third Inter-Agency Sedimentation Conference, Denver, Colorado March 22–26, Part 4*, pp. 100–113, U.S. Gov. Print. Off., Washington, D. C.

Emmett, W. W., and M. G. Wolman (2001), Effective discharge and gravel-bed rivers, *Earth Surf. Processes Landforms*, *26*, 1369–1380.

Engelund, F. (1981), Transport of bed load at high shear stress, Progress Report No. *53*, 31–35, Institute of Hydrodynamics and Hydraulic Engineering, ISVA, Tech. Univ. of Denmark.

Engelund, F., and J. Fredsøe (1976), A sediment transport model for straight alluvial channels, *Nord. Hydrol.*, *7*, 293–306.

Engelund, F., and E. Hansen (1967), *A Monograph on Sediment Transport in Alluvial Streams*, 65 pp., Teknisk Forlag, Copenhagen.

Evenson, E. B., and J. M. Clinch (1987), Debris transport mechanisms at active alpine glacier margins: Alaskan case studies, in *INQUA Till Symposium*, *Geol. Surv. of Finland Spec Pap. 3*, edited by R. Kujansuu and M. Saarnisto, pp. 111–136, Helsinki.

Everaert, W. (1991), Empirical relations for the sediment transport capacity of interrill flow, *Earth Surf. Processes Landforms*, *16*, 513–532.

Everaert, W. (1992), Mechanismen van intergeulerosie: Laboratorium experimenten, PhD thesis, Katholieke Universitet Leuven, Leuven, Belgium.

Farres, P. J. (1987), The dynamics of rainsplash erosion and the role of soil aggregate stability, *Catena*, *14*, 119–130.

Favis-Mortlock, D. T., J. Boardman, A. J. Parsons, and B. Lascelles (2000), Emergence and erosion: A model for rill initiation and development, *Hydrol. Processes*, *14*, 2173–2205.

Ferguson, R. I. (2003), The missing dimension: Effects of lateral variation on 1-D calculations of fluvial bedload transport, *Geomorphology*, *56*(1–2), 1–14.

Ferguson, R. I. (2008), Gravel-bed rivers at the reach scale, in *Gravel-Bed Rivers VI: From Process Understanding to River Restoration*, edited by H. Habersack, H. Piégay, and M. Rinaldi, pp. 33–60, Elsevier, Amsterdam.

Ferguson, R. I., K. L. Prestegaard, and P. J. Ashworth (1989), Influence of sand on hydraulics and gravel transport in a braided gravel bed river, *Water Resour. Res.*, *25*, 635–643, doi:10.1029/WR025i004p00635.

Ferguson, R. I., D. J. Bloomer, T. B. Hoey, and A. Werritty (2002), Mobility of river tracer pebbles over different timescales, *Water Resour. Res.*, *38*(5), 1045, doi:10.1029/2001WR000254.

Fernandez Luque, R., and R. van Beek (1976), Erosion and transport of bedload sediment, *J. Hydraul. Res.*, *14*, 127–144.

Ferro, V. (1998), Evaluating overland flow sediment transport capacity, *Hydrol. Processes*, *12*, 1895–1910.

Foster, G. R. (1982), Modeling the erosion process, in *Hydrologic Modeling of Small Watersheds*, *ASAE Monogr.*, vol. 5, edited by C. T. Haan, pp. 297–360, Am. Soc. Agric. Eng., St. Joseph, Mich.

Foster, G. R., and L. D. Meyer (1972), A closed form soil-erosion equation for upland areas, in *Sedimentation*, edited by H. W. Shen, pp. 12.1–12.19, Colorado State Univ., Fort Collins, Colo.

Fourrière, A., P. Claudin, and B. Andreotti (2010), Bedforms in a stream: Formation of ripples by primary linear instability and of dunes by non-linear pattern coarsening, *J. Fluid Mech.*, *649*, 287–328.

Francis, J. R. D. (1973), Experiments on the motion of solitary grains along the bed of a water-stream, *Trans. R. Soc. London, Ser. A*, *332*, 443–471.

Fredsoe, J. (1993), Modelling of non-cohesive sediment transport processes in the marine environment, *Coastal Eng.*, *21*, 71–103.

Frey, P., and M. Church (2011), Bedload: A granular phenomenon, *Earth Surf. Processes Landforms*, *36*, 58–69.

Fuller, I. C., and M. Marden (2010), Rapid channel response to variability in sediment supply: Cutting and filling of the Tarndale Fan, Waipaoa catchment, New Zealand, *Mar. Geol.*, *270*, 45–54.

Furbish, D. J., P. K. Haff, J. C. Roseberry, and M. W. Schmeeckle (2012), A probabilistic description of the bed load sediment flux: 1. Theory, *J. Geophys. Res.*, *117*, F03031, doi:10.1029/2012JF002352.

Galeone, D. G. (1996), Calibration of paired basins prior to streambank fencing of pasture land, *J. Environ. Qual.*, *28*, 1853–1863.

Gao, P. (2003), Mechanics of bedload transport in the saltation and sheetflow regimes, Unpublished PhD thesis, Geography, State Univ. of New York, Univ. of Buffalo, Buffalo, New York.

Gao, P. (2008), Transition between two bedload transport regimes: Saltation and sheet flow, *J. Hydraul. Eng.*, *134*, 340–349.

Gao, P. (2011), An equation for bedload transport capacities in gravel-bed rivers, *J. Hydrol.*, *402*, 297–305.

Gao, P. (2012), Validation and implications of an energy-based bedload transport equation, *Sedimentology*, *59*, 1926–1935.

Gares, P. A., R. G. D. Davidson-Arnott, B. O. Bauer, D. J. Sherman, R. W. G. Carter, D. W. T. Jackson, and K. F. Nordstrom (1996), Alongshore variations in aeolian sediment transport, Carrick Finn Strand, Ireland, *J. Coastal Res.*, *12*, 673–682.

Gilbert, G. K. (1877), Report on the geology of the Henry Mountains, Department of the Interior, U. S. Geographical and Geol. Surv. of Mountain Region, Washington, D. C, 170 pp.

Gilbert, G. K. (1914), *The Transportation of Debris by Running Water*, *U. S. Geol. Surv. Prof. Pap.*, vol. 86, pp. 263, U.S. Geol. Surv., Washington, D. C.

Gilbert, G. K. (1917), *Hydraulic Mining Débris in the Sierra Nevada*, *U. S. Geol. Surv. Prof. Pap.*, vol. 105, U.S. Geol. Surv., Washington, D. C.

Gillette, D. A. (1977), Fine particulate emissions due to wind erosion, *Trans. ASAE*, *20*(5), 890–897.

Gillette, D. A., and W. Chen (2001), Particle production and aeolian transport from a "supply-limited" source area in the Chihuahuan Desert, New Mexico, United States, *J. Geophys. Res.*, *106*, 5267–5278, doi:10.1029/2000JD900674.

Gillette, D. A., J. Adams, E. Endo, and D. Smith (1980), Threshold velocities for input of soil particles into the air by desert soils, *J. Geophys. Res.*, *85*, 5621–5630, doi:10.1029/JC085iC10p05621.

Gillette, D. A., E. Hardebeck, and J. Parker (1997), Large-scale variability of wind erosion mass flux rates at Owens Lake: 2. Role of roughness change, particle limitation, change of threshold friction velocity and the Owen effect, *J. Geophys. Res.*, *102*, 25,989–25,998, doi:10.1029/97JD00960.

Gillies, J. A. (2013), Fundamentals of aeolian sediment transport: Dust emissions and transport—Near surface, in *Treatise on Geomorphology*, edited by J. Shroder and N. Lancaster, pp. 43–63, Academic Press, San Diego, Calif.

Gillies, J. A., and N. Lancaster (2013), Large roughness element effects on sand transport, Oceano Dunes, California, *Earth Surf. Processes Landforms*, *38*, 785–792, doi:10.1002/esp.3317.

Gillies, J. A., W. G. Nickling, and G. McTainsh (1996), Dust concentrations and particle-size characteristics of an intense dust haze event: Inland Delta Region, Mali, West Africa, *Atmos. Environ.*, *30*(7), 1081–1090.

Gillies, J. A., W. G. Nickling, and J. King (2006), Aeolian sediment transport through large patches of roughness in the atmospheric inertial sublayer, *J. Geophys. Res.*, *111*, F02006, doi:10.1029/2005JF000434.

Glen, J. W. (1952), Experiments on the deformation of ice, *J. Glaciol.*, *2*, 111–114.

Gob, F., J. Bravard, and F. Petit (2010), The influence of sediment size, relative grain size and channel slope on initiation of sediment motion in boulder bed rivers. A lichenometric study, *Earth Surf. Processes Landforms*, *35*, 1535–1547.

Gomez, B. (1983), Temporal variations in bedload transport rates: The effect of progressive bed armouring, *Earth Surf. Processes Landforms*, *8*, 41–54.

Gomez, B. (1995), Bedload transport and changing grain size distribution, in *Changing River Channels*, edited by A. M. Gurnell and G. Petts, pp. 177–199, John Wiley, Chichester.

Gomez, B. (2006), The potential rate of bedload transport, *Proc. Natl. Acad. Sci. U.S.A.*, *103*, 17,170–17,173.

Gomez, B., and M. Church (1988), *A Catalogue of Equilibrium Bedload Transport Data for Coarse Sand Gravel-Bed Channels*, pp. 1–90, Department of Geography, Univ. of British Columbia, Vancouver.

Gomez, B., and M. Church (1989), An assessment of bed load sediment transport formulae for gravel bed rivers, *Water Resour. Res.*, *25*, 1161–1186, doi:10.1029/WR025i006p01161.

Gomez, B., R. L. Naff, and D. W. Hubbell (1989), Temporal variations in bedload transport rates associated with the migration of bedforms, *Earth Surf. Processes Landforms*, *14*, 135–156.

Gomez, B., B. J. Rosser, D. H. Peacock, D. M. Hicks, and J. A. Palmer (2001), Downstream fining in a rapidly aggrading gravel bed river, *Water Resour. Res.*, *37*, 1813–1823, doi:10.1029/2001WR900007.

Govers, G. (1985), Selectivity and transport capacity of thin flows in relation to rill generation, *Catena*, *12*, 35–49.

Govers, G. (1990), Empirical relationships for the transport capacity of overland flow, in *Erosion, Transport and Deposition Processes, IAHS-AISH Publ.*, vol. 189, edited by D. E. Walling, A. Yair, and S. Berkowicz, pp. 45–63, IAHS, Wallingford.

Govers, G. (1991), Spatial and temporal variations in splash detachment: A field study, *Catena Suppl.*, *20*, 15–24.

Govers, G. (1992), Evaluation of transport capacity formulae for overland flow, in *Overland Flow: Hydraulics and Erosion Mechanics*, edited by A. J. Parsons and A. D. Abrahams, pp. 243–273, UCL Press, London.

Graf, W. H. (1971), *Hydraulics of Sediment Transport*, McGraw-Hill Book Company, New York.

Graf, W. H. (1991), Flow resistance over a gravel bed: Its consequences on initial sediment movement, in *Fluvial Hydraulics in Mountain Regions*, edited by A. Armanini and G. Di Silvio, pp. 17–32, Springer, Berlin.

Graf, W. H., and M. Cellino (2002), Suspension flows in open channels: Experimental study, *J. Hydraul. Res.*, *40*, 435–447.

Graf, W. H., and L. Suszka (1987), Sediment transport in steep channels, *J. Hydrosci. Hydraul. Eng.*, *5*, 11–26.

Grant, S. B., and I. Marusic (2011), Crossing turbulent boundaries: Interfacial flux in environmental flows, *Environ. Sci. Technol.*, *45*(17), 7107–7113, doi:10.1021/es201778s.

Greeley, R., D. G. Blumberg, and S. H. Williams (1996), Field measurements of the flux and speed of wind-blown sand, *Sedimentology*, *43*, 41–52.

Green, M. O., and G. Coco (2014), Review of wave-driven sediment resuspension and transport in estuaries, *Rev. Geophys.*, *52*, 77–117, doi:10.1002/2013RG000437.

Hacking, I. (1983), *Representing and Intervening, Introductory Topics in the Philosophy of Natural Science*, Cambridge Univ. Press, Cambridge.

Hagen, J. O. (1987), Glacier surge at Usherbreen, Svalbard, *Polar Res.*, *5*, 239–252.

Hairsine, P. B., and C. W. Rose (1992a), Modeling water erosion due to overland-flow using physical principles: 1. Sheet flow, *Water Resour. Res.*, *28*(1), 237–243, doi:10.1029/91WR02380.

Hairsine, P. B., and C. W. Rose (1992b), Modeling water erosion due to overland-flow using physical principles: 2. Rill flow, *Water Resour. Res.*, *28*(1), 245–250, doi:10.1029/91WR02381.

Hardisty, J. (1993), Monitoring and modelling sediment transport at turbulent frequencies, in *Turbulence: Perspectives on Fluid Flow and Sediment Transport*, edited by N. J. Clifford, J. R. French, and J. Hardisty, pp. 35–59, John Wiley, Chichester.

Hassan, M. A., M. Church, and A. P. Schick (1991), Distance of movement of coarse particles in gravel bed streams, *Water Resour. Res.*, *27*, 503–511, doi:10.1029/90WR02762.

Hassan, M. A., R. Egozi, and G. Parker (2006), Experiments on the effect of hydrograph characteristics on vertical grain sorting in gravel bed rivers, *Water Resour. Res.*, *42*, W09408, doi:10.1029/2005WR004707.

Hayes, S. K. (1999), Low-flow sediment transport on the Pasig-Potrero alluvial fan, Mount Pinatubo, Philippines, M.S., Department of Geol. Sci., Univ. of Washington, Seattle.

Hersen, P., S. Douady, and B. Andreotti (2002), Relevant length scale of barchan dunes, *Phys. Rev. Lett.*, *89*(26), 264,301–264,304.

Hessel, R., and V. Jetten (2007), Suitability of transport equations in modelling soil erosion for a small Loess Plateau catchment, *Eng. Geol.*, *91*, 56–71.

Hey, R. D., and C. R. Thorne (1986), Stable channels with mobile gravel beds, *J. Hydraul. Eng.*, *112*, 671–689.

Hicks, D. M., and B. Gomez (2003), Sediment transport, in *Tools in Fluvial Geomorphology*, edited by G. M. Kondolf and H. Piégay, pp. 425–462, John Wiley, Chichester.

Hill, K. M., L. DellAngelo, and M. M. Meerschaert (2010), Heavy-tailed travel distance in gravel bed transport: An exploratory enquiry, *J. Geophys.*, *115*, F00A14, doi:10.1029/2009JF001276.

Holden, J. (Ed.) (2008), *An Introduction to Physical Geography and the Environment*, Prentice Hall, Harlow.

Holmes, A. (1915), Review: The Transportation of Débris by Running Water. By G.K. Gilbert. U.S. Geol. Surv. Prof. Pap. 86. pp. 263. 1914, *Geol. Mag. (Decade VI)*, *2*, 132–134.

Hooke, R. L. (1975), Distribution of sediment transport and shear stress in a meander bend, *J. Geol.*, *83*, 543–565.

Hopkins, W. (1844), On the transport of erratic blocks, *Trans. Cambridge Philos. Soc.*, *VIII*, 220–240.

Horikawa, K. (1978), *Coastal Engineering*, 402 pp., Univ. of Tokyo Press, Tokyo.

Horikawa, K. (Ed.) (1988), *Nearshore Dynamics and Coastal Processes*, 522 pp., Univ. of Tokyo Press, Tokyo.

Hsu, S. M., L. B. Chiou, G. F. Lin, C. H. Chao, H. Y. Wen, and C. Y. Ku (2010), Applications of simulation technique on debris-flow hazard zone delineation: A case study in Hualien County, Taiwan, *Nat. Hazards Earth Syst. Sci.*, *10*, 535–545.

Hu, C., and Y. Hui (1996), Bedload transport. I: Mechanical characteristics, *J. Hydraul. Eng.*, *122*, 245–254.

Huang, C., L. K. Wells, and L. D. Norton (1999), Sediment transport capacity and erosion processes: Model concepts and reality, *Earth Surf. Processes Landforms*, *24*, 503–516.

Hughes, M. G., T. Aagaard, T. E. Baldock, and H. E. Power (2014), Spectral signatures for swash on reflective, intermediate and dissipative beaches, *Mar. Geol.*, *355*, 88–97.

Hungr, O., G. C. Morgan, and R. Kellerhals (1984), Quantitative analysis of debris torrent hazards for design of remedial measures, *Can. Geotech. J.*, *21*, 663–677.

Hungr, O., S. McDougall, and M. Bovis (2005), Entrainment of material by debris flows, in *Debris-Flow Hazards and Related Phenomena*, edited by M. Jakob and O. Hungr, pp. 135–158, Springer, Berlin, doi:10.1007/b138657.

Hunter, L. E., R. D. Powell, and D. E. Lawson (1996), Flux of debris transported by ice at three Alaskan tidewater glaciers, *J. Glaciol.*, *42*(140), 123–135.

Huntley, D. A., R. T. Guza, and A. J. Bowen (1977), A universal form for shoreline run-up spectra?, *J. Geophys. Res.*, *82*, 2577–2581, doi:10.1029/JC082i018p02577.

Ikeya, H. (1981), A method of designation for area in danger of debris flow, in *Erosion and Sediment Transport in Pacific Rim Steeplands*, *IAHS Publ.*, vol. 132, edited by T. R. H. Davies and A. J. Pearce, pp. 576–587, IAHS Wallingford.

Inman, D. L., and R. A. Bagnold (1963), Beach and nearshore processes: Part II. Littoral processes, in *The Sea: Ideas and Observations on Progress in the Study of the Sea*, vol. 3, edited by M. N. Hill, pp. 529–553, Wiley-Interscience, London.

Iseya, F., and H. Ikeda (1987), Pulsations in bedload transport rates induced by a longitudinal sediment sorting: A flume study using sand and gravel mixtures, *Geogr. Ann.*, *69A*, 15–27.

Iverson, N. R. (2000), Regelation of ice through debris at glacier beds: Implications for sediment transport, *Geology*, *21*, 559–562.

Iverson, R. M. (1997), The physics of debris flow, *Rev. Geophys.*, *35*, 245–296, doi:10.1029/97RG00426.

Iverson, R. M., and R. P. Denlinger (1987), The physics of debris flows—A conceptual assessment, in *Erosion and Sedimentation in the Pacific Rim (Proceedings of the Corvallis Symposium, August, 1987)*, *IAHS Publ.*, vol. 165, edited by R. L. Beschta et al., pp. 155–166, IAHS, Wallingford.

Iverson, R. M., S. P. Schilling, and J. W. Vallance (1998), Objective delineation of lahar-inundation hazard zones, *Geol. Soc. Am. Bull.*, *110*(8), 972–984.

Jackson, N. L., D. J. Sherman, P. A. Hesp, A. H. F. Klein, F. Ballasteros, and K. F. Nordstrom (2006), Small-scale spatial variations in aeolian sediment transport on a fine-sand beach, *J. Coastal Res.*, *39*, 379–383.

Jackson, R. G. (1976), Sedimentological and fluid-dynamic implications of the turbulent bursting phenomenon in geophysical flows, *J. Fluid Mech.*, *77*, 531–560.

Jackson, W. L., and R. L. Beschta (1982), A model of two-phase bedload transport in an Oregon coast range stream, *Earth Surf. Processes Landforms*, *7*, 517–527.

Jakob, M., and H. Weatherly (2003), A hydroclimatic threshold for shallow landslide initiation at the North Shore Mountains of Vancouver, BC, *Geomorphology*, *57*, 137–156.

Jakob, M., M. Bovis, and M. Oden (2005), The significance of channel recharge rates for estimating debris-flow magnitude and frequency, *Earth Surf. Processes Landforms*, *30*, 755–766.

Johnson, J. W. (1943), Laboratory investigations on bedload transportation and bed roughness: A compilation of published and unpublished data, U.S. Soil Consev. Serv., SCS-TP-50, Washington, D. C.

Julien, P. Y. (1987), Sediment transport capacity of sheet and rill flow - application of unit stream power theory – comment, *Water Resour. Res.*, *23*, 1700.

Julien, P. Y. (1998), *Erosion and Sedimentation*, Cambridge Univ. Press, Cambridge.

Julien, P. Y., and D. B. Simons (1985), Sediment transport capacity of overland flow, *Trans. ASAE*, *28*, 755–762.

Kamphuis, J. W. (1991), Alongshore sediment transport rate, *J. Waterw. Port Coastal Ocean Eng.*, *117*, 624–640.

Kawamura, R. (1951), Study on sand movement by wind, *Reports of the Physical Science Research Institute of Tokyo University*, vol. 5, pp. 95–112.

Kennedy, R. G. (1895), The prevention of silting in irrigation canals, *Min. Proc. Inst. Civil Eng.*, *119*, 281–290.

Kilinc, M., and E. V. Richardson (1973), *Mechanics of Soil Erosion from Overland Flow Generated by Simulated Rainfall*, *Hydrol. Pap.*, vol. 63, Colorado State Univ., Boulder, Colo.

Kind, R. J. (1976), A critical examination of the requirements for model simulation of wind-induced erosion/deposition phenomena such as snow drifting, *Atmos. Environ.*, *10*, 219–227.

Kirkby, M. J. (1993), Long-term interactions between networks and hillslopes, in *Channel Network Hydrology*, edited by K. J. Beven and M. J. Kirkby, pp. 254–293, John Wiley, Chichester.

Kirkby, M. J. (1997), TOPMODEL: A personal view, *Hydrol. Processes*, *11*, 1087–1097.

Kirkby, M. J., and I. Statham (1975), Surface stone movement and scree formation, *J. Geol.*, *83*, 349–362.

Kjelgaard, J., B. Sharratt, I. Sundram, B. Lamb, C. Claiborn, K. Saxton, and D. Chandler (2004a), PM10 emission from agricultural soils on the Columbia Plateau: Comparison of dynamic and time-integrated field-scale measurements and entrainment mechanisms, *Agric. Forest Meteorol.*, *125*(3–4), 259–277.

Kjelgaard, J. F., D. G. Chandler, and K. E. Saxton (2004b), Evidence for direct suspension of loessial soils on the Columbia Plateau, *Earth Surf. Processes Landforms*, *29*(2), 221–236, doi:10.1002/esp.1028.

Knight, P. G., R. I. Waller, C. J. Patterson, A. P. Jones, and Z. P. Robinson (2002), Discharge of debris from ice at the margin of the Greenland ice sheet, *J. Glaciol.*, *48*(161), 192–198.

Knighton, A. D. (1989), River adjustment to changes in sediment load: The effects of tin mining on the Ringarooma River, Tasmania, *Earth Surf. Processes Landforms*, *14*, 333–359.

Knighton, A. D. (1998), *Fluvial Forms and Processes: A New Perspective*, 2nd ed., 383 pp., Arnold, London.

Kok, J. F. (2011), A scaling theory for the size distribution of emitted dust aerosols suggests climate models underestimate the size of the global dust cycle, *Proc. Natl. Acad. Sci. U.S.A.*, *108*(3), 1016–1021.

Komar, P. D. (1980), K-type coefficients in the littoral drift system: Variable or constant, *Mar. Geol.*, *34*, M99–M101.

Komar, P. D. (1999), *Beach Processes and Sedimentation*, 2nd ed., 544 pp., Prentice Hall, Harlow.

Krecic, M. R., and D. M. Hanes (1996), An analysis of particle saltation dynamics, in *Coastal Engineering*, edited by B. L. Edge, pp. 3846–3859, ASCE, New York.

Kronvang, B., and A. J. Bruhn (1996), Choice of sampling strategy and estimation method for calculating nitrogen and phosphorus transport in small lowland streams, *Hydrol. Processes*, *10*, 1483–1501.

Kronvang, B., J. Audet, A. Baattrup-Pedersen, H. S. Jensen, and S. E. Larsen (2012), Phosphorus load to surface water from bank erosion in a Danish lowland river basin, *J. Environ. Qual.*, *41*, 304–313.

Kuhnle, R. A. (1992), Bed load transport during rising and falling stages on two small streams, *Earth Surf. Processes Landforms*, *17*, 191–197.

Lacey, G. (1930), Stable channels in alluvium, *Min. Proc. Inst. Civil Eng.*, *229*, 259–292.

Lajeunesse, E., L. Malverti, and F. Charru (2010), Bed load transport in turbulent flow at the grain scale: Experiments and modeling, *J. Geophys. Res.*, *115*, F04001, doi:10.1029/02009JF001628.

Lamb, M. P., W. E. Dietrich, and J. G. Venditti (2008), Is the critical Shields stress for incipient sediment motion dependent on channel-bed slope?, *J. Geophys. Res.*, *113*, F02008, doi:10.1029/02007JF000831.

Laronne, J. B., and M. A. Carson (1976), Interrelationships between bed morphology and bed-material transport for a small, gravel-bed channel, *Sedimentology*, *23*, 67–85.

Laronne, J. B., and I. Reid (1993), Very high rates of bed load sediment transport in ephemeral desert rivers, *Nature*, *366*, 148–150.

Laronne, J. B., I. Reid, Y. Yitshak, and L. E. Frostick (1994), The non-layering of gravel streambeds under ephemeral flood regimes, *J. Hydrol.*, *159*, 353–363.

Latour, B. (1987), *Science in Action: How to Follow Scientists and Engineers Through Society*, Harvard Univ. Press, Cambridge, Mass.

Laursen, E. M. (1958), The total sediment load of streams, *J. Hydraul. Div., Am. Soc. Civ. Eng.*, *84*(HY1), 1–36.

Lechalas, C. (1871), Note sur les rivières à fond de sable, *Annales des Ponts et Chaussées, Mémoires et Documents, 5ème série*, *1*, 381–431.

Lee, H., and I. Hsu (1994), Investigation of saltating particle motions, *J. Hydraul. Eng.*, *120*, 831–845.

Lee, H., Y. Chen, J. You, and Y. Lin (2000), Investigations of continuous bed load saltating process, *J. Hydraul. Eng.*, *126*, 691–700.

Lee, K. T., Y. L. Liu, and K. H. Cheng (2004), Experimental investigation of bed load transport processes under unsteady flow conditions, *Hydrol. Processes*, *18*, 2439–2454.

Leopold, L. B. (1980), Techniques and interpretation: The sediment studies of GK Gilbert, *Geol. Soc. Am., Spec. Pap.*, *183*, 125–128.

Leopold, L. B., M. G. Wolman, and J. P. Miller (1964), *Fluvial Processes in Geomorphology*, 522 pp., WH Freeman and Company, San Francisco, Calif.

Leslie, J. (1823), *Elements of Natural Philosophy. Volume First, Including Mechanics and Hydrostatics*, W & C Tait, Edinburgh.

Lettau, K., and H. H. Lettau (1978), Experimental and micro-meteorological field studies on dune migration, in *Exploring the World's Driest Climate*, edited by K. Lettau and H. H. Lettau, pp. 110–147, Univ. of Wisconsin-Madison, Inst. for Environ. Studies, Madison, Wis.

Li, B., H. M. Granja, E. J. Farrell, J. T. Ellis, and D. J. Sherman (2009), Aeolian saltation at Esposende Beach, Portugal, *J. Coastal Res.*, *S156*, 327–331.

Li, B., D. J. Sherman, E. J. Farrell, and J. T. Ellis (2010), Variability of the apparent von Kármán parameter during aeolian saltation, *Geophys. Res. Lett.*, *37*, L15404, doi:10.1029/2010GL044068.

Lien, H. P., and F. W. Tsai (2003), Sediment concentration distribution of debris flow, *J. Hydraul. Eng., ASCE*, *129*(12), 995–1000.

Lisle, T. E. (1995), Particle size variations between bed load and bed material in natural gravel bed channels, *Water Resour. Res.*, *31*, 1107–1118, doi:10.1029/94WR02526.

Lisle, T. E. (2008), The evolution of sediment waves influenced by varying transport capacity in heterogeneous rivers, in *Gravel-Bed Rivers VI: From Process Understanding to River Restoration*, edited by H. Habersack, H. Piégay, and M. Rinaldi, pp. 443–469, Elsevier, Amsterdam.

Lisle, T. E., and M. Church (2002), Sediment transport-storage relations for degrading, gravel bed channels, *Water Resour. Res.*, *38*(11), 1219, doi:10.1029/2001WR001086.

Lisle, T. E., and M. A. Madej (1992), Spatial variation in armouring in a channel with high sediment supply, in *Dynamics of Gravel-Bed Rivers*, edited by P. Billi et al., pp. 277–296, John Wiley, Chichester.

Lisle, T. E., and B. Smith (2003), Dynamic transport capacity in gravel-bed river systems, in *International Workshop for Source to Sink Sedimentary Dynamics in Catchment Scale*, edited by T. Araya, M. Kuroki, and T. Marutani, Organizing Committee of the International Workshop for Sedimentary Dynamics, Sapporo, Japan.

Lliboutry, L. (1993), Internal melting and ice accretion at the bottom of temperate glaciers, *J. Glaciol.*, *39*, 50–64.

Long, E. J., G. K. Hargrave, J. R. Cooper, B. G. B. Kitchener, A. J. Parsons, C. Hewett, and J. Wainwright (2014), Experimental investigation into the impact of a liquid droplet onto a granular bed using 3D, time-resolved, particle tracking, *Phys. Rev. E*, *89*, 032201, doi:10.1103/PhysRevE.89.032201.

Mackin, J. H. (1948), Concept of the graded river, *Geol. Soc. Am. Bull.*, *59*, 463–512.

Mao, L. (2012), The effect of hydrographs on bed load transport and bed sediment spatial arrangement, *J. Geophys. Res.*, *117*, F03024, doi:10.1029/2012JF002428.

Mao, L., G. P. Uyttendaele, A. Iroumé, and M. A. Lenzi (2008), Field based analysis of sediment entrainment in two high gradient streams located in Alpine and Andine environments, *Geomorphology*, *93*, 368–383.

Marchi, L., M. Arattano, and A. M. Deganutti (2002), Ten years of debris flows monitoring in the Moscardo Torrent (Italian Alps), *Geomorphology*, *46*, 1–17.

Marcus, W. A. (1989), Lag-time routing of suspended sediment concentrations during unsteady flow, *Geol. Soc. Am. Bull.*, *101*, 644–651.

Marti, C., and G. R. Bezzola (2006), Bed load transport in braided gravel-bed rivers, in *Braided Rivers: Process, Deposits, Ecology and Management*, edited by G. H. Sambrook Smith et al., pp. 199–216, Oxford, Blackwell.

Martin, Y. (2003), Evaluation of bed load transport formulae using field evidence from the Vedder River, British Columbia, *Geomorphology*, *53*, 75–95.

McDowell, D. M. (1989), A general formula for estimation of the rate of transport of bed load by water, *J. Hydraul. Res.*, *27*, 355–361.

McEwan, I. K. (1991), The physics of sand transport by wind, Unpublished PhD thesis, Univ. of Aberdeen, Aberdeen, 121 pp.

Melanson, A., T. Bell, and L. Tarasov (2013), Numerical modelling of subglacial erosion and sediment transport and its application to the North American ice sheets over the Last Glacial cycle, *Quat. Sci. Rev.*, *68*, 154–174.

Merton, G. H., M. A. Nearing, and A. L. O. Borges (2001), Effect of sediment load on soil detachment and deposition in rills, *Soil Sci. Soc. Am. J.*, *65*, 861–868.

Meyer, L. D., and W. H. Wischmeier (1969), Mathematical simulation of the process of soil erosion by water, *Trans. ASAE*, *12*, 754–758, 762.

Meyer, L. D., G. R. Foster, and M. J. M. Römkens (1972), Source of soil eroded by water from upland slopes, in *Present and Prospective Technology for Predicting Sediment Yields and Sources, Proceedings of the Sediment-Yield Workshop, Oxford, Mississippi, November 1972*, pp. 177–189, USDA-ARS ARS-S-40, Washington, D. C.

Meyer, L. D., G. R. Foster, and S. Nikolov (1975), Effect of flow-rate and canopy on rill erosion, *Trans. ASAE*, *18*, 905–911.

Meyer-Peter, E., and R. Müller (1948), Formulas for bed load transport, *Proceedings of the International Association for Hydraulic Research, Third Annual Conference*, pp. 39–64, Stockholm, Sweden.

Meyer-Peter, E., H. Favre, and H. A. Einstein (1934), Neuere Versuchsresultate über den Geschiebetrieb, *Schweiz. Bauztg.*, *103*(13), 147–150.

Mil-Homens, J., R. Ranasinghe, R. van Thiel, J. S. M. de Vries, and M. J. F. Stive (2013), Re-evaluation and improvement of three commonly used bulk longshore sediment transport formulas, *Coastal Eng.*, *75*, 29–39.

Misra, R. K., and C. W. Rose (1990), Annual for use of program GUEST, Division of Australian Environmental Studies Report, Griffith Univ., Brisbane, Qld.

Molinas, A., and B. Wu (2001), Transport of sediment in large sand-bed rivers, *J. Hydraul. Res.*, *39*, 135–145.

Montgomery, D. R., M. S. Panfil, and S. K. Hayes (2000), Channel-bed mobility response to extreme sediment loading at Mount Pinatubo, *Geology*, *27*, 271–274.

Moore, I. D., and G. J. Burch (1986), Sediment transport capacity of sheet and rill flow: Application of unit stream power theory, *Water Resour. Res.*, *22*, 1350–1360, doi:10.1029/WR022i008p01350.

Moore, I. D., and G. J. Burch (1987), Reply [to "Comments on 'Sediment transport capacity of sheet and rill flow: Application of unit stream power theory' by I.D. Moore and G. J. Burch"], *Water Resour. Res.*, *23*, 1701–1702, doi:10.1029/WR023i008p01701.

Morfett, J. C. (1990), A virtual power function for estimating the along shore transport of sediment by waves, *Coastal Eng.*, *14*, 439–456.

Morgan, R. P. C., J. N. Quinton, R. E. Smith, G. Govers, J. W. A. Poesen, K. Auerswald, G. Chisci, D. Torri, and M. E. Styczen (1998), The European Soil Erosion Model (EUROSEM): A dynamic approach for predicting sediment transport from fields and small catchments, *Earth Surf. Processes Landforms*, *23*, 527–544.

Mueller, E. R., and J. Pitlick (2005), Morphologically based model of bed load transport capacity in a headwater stream, *J. Geophys. Res.*, *110*, F02016, doi:10.1029/2003JF000117.

Mueller, E. R., J. Pitlick, and J. M. Nelson (2005), Variation in the reference Shields stress for bed load transport in gravel-bed streams and rivers, *Water Resour. Res.*, *41*, W04006, doi:10.1029/2004WR003692.

Musgrave, G. W. (1947), The quantitative evaluation of factors in water erosion: A first approximation, *J. Soil Water Conserv.*, *2*, 133–138, 170.

Namikas, S. L. (2003), Field measurement and numerical modeling of aeolian mass flux distributions on a sandy beach, *Sedimentology*, *50*, 303–326.

Nearing, M. A., G. R. Foster, L. J. Lane, and S. C. Finkner (1989), A process based soil erosion model for USDA water erosion prediction project technology, *Trans. ASAE*, *32*, 1587–1593.

Niño, Y., and M. Garcia (1998), Experiments on saltation of sand in water, *J. Hydraul. Eng., ASCE*, *124*, 1014–1025.

Niño, Y., M. García, and L. Ayala (1994), Gravel saltation: 1. Experiments, *Water Resour. Res.*, *30*, 1907–1914, doi:10.1029/94WR00533.

O'Brien, M. P., and B. D. Rindlaub (1936), The transportation of sand by wind, *Civil Eng.*, *6*, 325–327.

Ono, D., P. Kiddoo, C. Howard, G. Davis, and K. Richmond (2011), Application of a combined measurement and modeling method to quantify windblown dust emissions from the exposed playa at Mono Lake, California, *J. Air Waste Manage. Assoc.*, *61*, 1036–1054.

Oostwoud Wijdenes, D. J., and P. Ergenzinger (1998), Erosion and sediment transport on steep marly hillslopes, Draix, Haute-Provence, France: An experimental field study, *Catena*, *33*, 179–200.

Orford, J., and E. Anthony (2013), Morphodynamic systems—Coastal gravel systems, in *Treatise on Geomorphology*, vol. 10, edited by J. Shroder Jr. and D. Sherman, pp. 245–266, Academic Press, San Diego, Calif.

Orlovsky, L., N. Orlovsky, and A. Durdyev (2005), Dust storms in Turkmenistan, *J. Arid Environ.*, *60*, 83–97.

Owen, P. R. (1964), Saltation of uniform grains in air, *J. Fluid Mech.*, *20*, 225–242.

Özhan, E. (1982), Laboratory study of breaker type effects on longshore sand transport, in *Mechanics of Sediment Transport*, *Proceedings of Euromech*, vol. 156, edited by B. M. Sumer and A. Müller, pp. 265–274, AA Balkema, Rotterdam.

Papa, M., S. Egashira, and T. Itoh (2004), Critical conditions of bed sediment entrainment due to debris flow, *Nat. Hazards and Earth Syst. Sci.*, *4*, 469–474.

Paphitis, D., and M. B. Collins (2005), Sand grain threshold, in relation to bed "stress history": An experimental study, *Sedimentology*, *52*(4), 827–838.

Parker, C., N. J. Clifford, and C. R. Thorne (2011), Understanding the influence of slope on the threshold of coarse grain motion: Revisiting critical stream power, *Geomorphology*, *126*(1–2), 51–65.

Parker, G. (2006), Transport of gravel and sediment mixtures, in *Sedimentation Engineering: Processes, Measurements, Modeling, and Practice*, edited by M. H. Garcia, pp. 165–251, Am. Soc. of Civil Eng., Reston, Va.

Parker, G., and P. C. Klingeman (1982), On why gravel bed streams are paved, *Water Resour. Res.*, *18*, 1409–1423, doi:10.1029/WR018i005p01409.

Parker, G., and C. M. Toro-Escobar (2002), Equal mobility of gravel in streams: The remains of the day, *Water Resour. Res.*, *38*(11), 1264, doi:10.1029/2001WR000669.

Parker, G., P. C. Klingeman, and D. L. McLean (1982), Bedload and size distribution in paved gravel-bed streams, *J. Hydraul. Div., Am. Soc. Civ. Eng.*, *108*(HY4), 544–571.

Parsons, A. J., J. Wainwright, and A. D. Abrahams (1993), Tracing sediment movement in interrill overland-flow on a semiarid grassland hillslope using magnetic-susceptibility, *Earth Surf. Processes Landforms*, *18*(8), 721–732.

Parsons, A. J., J. Wainwright, D. M. Powell, J. Kaduk, and R. E. Brazier (2004), A conceptual model for understanding and predicting erosion by water, *Earth Surf. Processes Landforms*, *29*, 1293–1302.

Parsons, A. J., R. E. Brazier, J. Wainwright, and D. M. Powell (2006), Scale relationships in hillslope runoff and erosion, *Earth Surf. Processes Landforms*, *31*, 1384–1393.

Parsons, A. J., J. Wainwright, T. Furukawa, and Y. Onda (2010), Using sediment travel distance to estimate medium-term erosion rates: A sixteen-year record, *Earth Surf. Processes Landforms*, *35*, 1694–1700.

Parsons, A. J., J. Cooper, and J. Wainwright (2015), What is suspended sediment?, *Earth Surf. Processes Landforms*, doi:10.1002/esp.3730, in press.

Parteli, E. J. R., O. Durán, and H. J. Herrmann (2007a), Minimal size of a barchan dune, *Phys. Rev. E*, *75*, 011301.

Parteli, E. J. R., O. Durán, and H. J. Herrmann (2007b), Reply to "Comment on 'Minimal size of a barchan dune'", *Phys. Rev. E*, *76*, 063302.

Pender, G., T. B. Hoey, C. Fuller, and I. K. McEwan (2001), Selective bedload transport during the degradation of a well sorted graded sediment bed, *J. Hydraul. Res.*, *39*, 269–277.

Phillips, J. D. (2013), Geomorphic responses to changes in instream flows: The flow-channel fitness model, *River Res. Appl.*, *29*, 1175–1194.

Pierson, T. C., and J. E. Costa (1987), A rheologic classification of subaerial sediment-water flows, in *Debris Flows/Avalanches: Process, Recognition, and Mitigation*, edited by J. E. Costa and G. F. Wieczorek, pp. 1–12, Geol. Soc. of Am., Boulder, Colo.

Poesen, J. (1985), An improved splash transport model, *Z. Geomorphol.*, *29*, 193–211.

Poesen, J., and D. Torri (1988), The effect of cup size on splash detachment and transport measurements. Part I: Field measurements, *Catena Suppl.*, *12*, 113–126.

Polyakov, V. O., and M. A. Nearing (2003), Sediment transport in rill flow under deposition and detachment conditions, *Catena*, *51*, 33–43.

Powell, D. M., I. Reid, and J. B. Laronne (1999), Hydraulic interpretation of cross-stream variations in bedload transport, *J. Hydraul. Eng.*, *125*, 1243–1252.

Powell, D. M., I. Reid, and J. B. Laronne (2001), Evolution of bedload grain-size distribution with increasing flow strength and the effect of flow duration on the caliber of bedload sediment yield in ephemeral gravel-bed rivers, *Water Resour. Res.*, *34*, 1463–1474, doi:10.1029/2000WR900342.

Prancevic, J. P., M. P. Lamb, and B. M. Fuller (2014), Incipient sediment motion across the river to debris-flow transition, *Geology*, *42*(3), 191–194, doi:10.1130/G34927.1.

Prosser, I. P., and P. Rustomji (2000), Sediment transport capacity relations for overland flow, *Prog. Phys. Geogr.*, *24*, 179–193.

Pye, K. (1987), *Aeolian Dust and Dust Deposits*, Academic Press, London.

Quinton, J. N. (1997), Reducing predictive uncertainty in model simulations: A comparison of two methods using the European Soil Erosion Model (EUROSEM), *Catena*, *30*, 101–117.

Radice, A., F. Ballio, and V. Nikora (2009), On statistical properties of bed load sediment concentration, *Water Resour. Res.*, *45*, W06501, doi:10.1029/2008WR007192.

Radice, A., V. Nikora, J. Campagnol, and F. Ballio (2013), Active interactions between turbulence and bed load: Conceptual picture and experimental evidence, *Water Resour. Res.*, *49*, 90–99, doi:10.1029/2012WR012255.

Raupach, M. R., D. E. Hughes, and H. A. Cleugh (2006), Momentum absorption in rough-wall boundary layers with sparse roughness elements in random and clustered distributions, *Boundary Layer Meteorol.*, *120*, 201–218, doi:10.1007/s10546-10006-19058-10544.

Recking, A. (2006), Etude expérimentale de l'influence du tri granulométrique sur le transport solide par charriage, Unpublished PhD thesis, Institut National des Sciences Appliquées de Lyon. [Available at http://theses.insa-lyon.fr/publication/2006ISAL0113/these.pdf, last accessed 8th January 2014.]

Recking, A. (2009), Theoretical development on the effects of changing flow hydraulics on incipient bed load motion, *Water Resour. Res.*, *45*, W04401, doi:10.1029/2008WR006826.

Reid, I., L. E. Frostick, and J. T. Layman (1985), The incidence and nature of bed load transport during flood flows in coarse-grained alluvial channels, *Earth Surf. Processes Landforms*, *10*, 33–44.

Reid, I., J. B. Laronne, and D. M. Powell (1995), The Nahal Yatir bedload database: Sediment dynamics in a gravel-bed ephemeral stream, *Earth Surf. Processes Landforms*, *20*, 845–857.

Reid, I., D. M. Powell, and J. B. Laronne (1996), Prediction of bedload transport by desert flash floods, *J. Hydraul. Eng.*, *122*, 170–173.

Reid, L. M., and T. Dunne (1996), *Rapid Evaluation of Sediment Budgets*, Catena Verlag, Reiskirchen.

Rice, C. T., and B. N. Wilson (1990), Analysis of dynamic variables of rill flow, ASAE Paper No. 902011 (Presentation at the 1990 International Summer Meeting, Columbus, Ohio, June 24–27, 1990). ASAE, St Joseph, Mich. 34 pp.

Richards, K. S. (1990), "Real" geomorphology, *Earth Surf. Processes Landforms*, *15*, 195–197.

Richards, K. S., and N. J. Clifford (1991), Fluvial geomorphology—Structured beds in gravelly rivers, *Prog. Phys. Geogr.*, *15*, 407–422.

Rickenmann, D. (1990), Debris flows 1987 in Switzerland: Modelling and sediment transport, in *Hydrology in Mountainous Regions II*, IAHS Publ., vol. 194, edited by R. O. Sinniger and M. Monbaron, pp. 371–378, IAHS, Wallingford.

Rickenmann, D. (1991), Hyperconcentrated flow and sediment transport at steep slopes, *J. Hydraul. Eng.*, *125*, 117–125.

Rickenmann, D. (2012), Alluvial steep channels: Flow resistance, bedload transport and transition to debris flows, in *Gravel Bed Rivers: Processes, Tools, Environment*, edited by M. Church, P. Biron, and A. Roy, pp. 386–397, John Wiley, Chichester.

Rickenmann, D., and A. Koschni (2010), Sediment loads due to fluvial transport and debris flows during the 2005 flood events in Switzerland, *Hydrol. Processes*, *24*, 993–1007, doi:10.1002/hyp.7536.

Robert, A. (2003), *River Processes: An Introduction to Fluvial Dynamics*, Arnold, London.

Rose, C. W., K. J. Coughlan, and B. Fenti (1998), Griggith University Erosion System Template (GUEST), in *Modelling Soil Erosion by Water*, *NATO ASI Series I*, vol. 55, edited by J. Boardman and D. Favis-Mortlock, pp. 399–412, Springer, Berlin.

Rouse, H. (1939), An analysis of sediment transportation in the light of fluid turbulence, *Soil Conserv. Serv. Tech. Pap. 25*, Washington, D. C., July 1939 (mimeographed).

Ryan, S. E., L. S. Porth, and C. A. Troendle (2002), Defining phases of bedload transport using piecewise regression, *Earth Surf. Processes Landforms*, *27*, 971–990.

Sander, G. C., J.-Y. Parlange, D. A. Barry, M. B. Parlange, and W. L. Hogarth (2007), Limitation of the transport capacity approach in sediment transport modeling, *Water Resour. Res.*, *43*, W02403, doi:10.1029/2006WR005177.

Sarre, R. D. (1988), Evaluation of aeolian sand transport equations using intertidal zone measurements, Saunton Sands, England, *Sedimentology*, *35*, 671–700.

Sauermann, G., K. Kroy, and H. J. Herrmann (2001), Continuum saltation model for sand dunes, *Phys. Rev. E*, *64*, 031305.

Saville, T., Jr. (1950), Model study of sand transport along an infinitely long straight beach, *AGU*, *31*, 555–565.

Sayer, A. (2000), *Realism and Social Science*, Sage, London.

Sayre, N. F. (2008), The genesis, history, and limits of carrying capacity, *Ann. Assoc. Am. Geogr.*, *98*, 120–134.

Scheidl, C., and D. Rickenmann (2010), Empirical prediction of debris-flow mobility and deposition on fans, *Earth Surf. Processes Landforms*, *35*, 157–173.

Scheidl, C., D. Rickenmann, and M. Chiari (2008), The use of airborne LiDAR data for the analysis of debris-flow events in Switzerland, *Nat. Hazards Earth Syst. Sci.*, *8*, 1113–1127.

Schiettecatte, W., K. Verbist, and D. Gabriels (2008), Assessment of detachment and sediment transport capacity of runoff by field experiments on a silt loam soil, *Earth Surf. Processes Landforms*, *33*, 1302–1314.

Sear, D. A. (1996), Sediment transport processes in pool-riffle sequences, *Earth Surf. Processes Landforms*, *21*, 241–262.

Sekine, M., and H. Kikkawa (1992), Mechanics of saltating grains. II, *J. Hydraul. Eng.*, *118*, 536–558.

Shao, Y. (2000), *Physics and Modelling of Wind Erosion*, 393 pp., Kluwer Acad., Dordrecht.

Shao, Y. (2004), Simplification of a dust emission scheme and comparison with data, *J. Geophys. Res.*, *109*, D10202, doi:10.1029/2003JD004372.

Shao, Y., and M. R. Raupach (1992), The overshoot and equilibration of saltation, *J. Geophys. Res.*, *97*(D18), 20,559–20,564, doi:10.1029/92JD02011.

Sherman, D. J., and E. J. Farrell (2008), Aerodynamic roughness lengths over movable beds: Comparison of wind tunnel and field data, *J. Geophys. Res.*, *113*, F02S08, doi:10.1029/2007JF000784.

Sherman, D. J., and B. Li (2012), Predicting aeolian sand transport rates: A reevaluation of models, *Aeolian Res.*, *3*(4), 371–378.

Sherman, D. J., D. W. T. Jackson, S. L. Namikas, and J. Wang (1998), Wind-blown sand on beaches: An evaluation of models, *Geomorphology*, *22*, 113–133.

Shields, A. (1936), Anwendung der Aehnlichkeitsmechanik und der Turbulenzforschung auf die Geschiebebewegung, *Mitteilungen der Preussischen Versuchsanstalt für Wasserbau und Schiffbau*, vol. 36.

Short, A. D. (Ed.) (1999a), *Handbook of Beach and Shoreface Morphodynamics*, John Wiley, Chichester.

Short, A. D. (1999b), Wave-dominated beaches, in *Handbook of Beach and Shoreface Morphodynamics*, edited by A. D. Short, pp. 173–203, John Wiley, Chichester.

Shu, A. P., and X. J. Fei (2008), Sediment transport capacity of hyperconcentrated flow, *Sci. China, Ser. G*, *51*, 961–975.

Shvidchenko, A. B., and G. Pender (2000), Flume study of the effect of relative depth on the incipient motion of coarse uniform sediments, *Water Resour. Res.*, *36*(2), 619–628, doi:10.1029/1999WR900312.

Sidorchuk, A. (1999), Dynamic and static models of gully erosion, *Catena, 37*, 401–414.

Simons, D. B., and F. Şentürk (1992), *Sediment Transport Technology: Water and Sediment Dynamics*, Water Resour. Publ., Littleton, Colo.

Smart, G. M. (1983), Sediment transport formula for steep channels, *J. Hydraul. Eng., 110*, 267–277.

Smart, G. M., and M. N. R. Jäggi (1983), Sediment transport on steep slopes, *Mitteilungen der Versuchsanstalt für Wasserbau, Hydrologie und Glaziologie*, vol. 64, Technische Hochschule (ETH), Zürich, Switzerland.

Smith, R. E., D. C. Goodrich, D. A. Woolhiser, and C. L. Unkrich (1995), KINEROS—A kinematic runoff and erosion model, in *Computer Models of Watershed Hydrology*, edited by V. P. Singh, pp. 697–732, Water Resour. Publ., Littleton Colo.

Smith, R. E., J. Quinton, D. C. Goodrich, and M. Nearing (2010), Soil-erosion models: Where do we really stand?, *Earth Surf. Processes Landforms, 35*, 1344–1348.

Sommerfeld, M. (2001), Validation of a stochastic Lagrangian modelling approach for inter-particle collisions in homogeneous isotropic turbulence, *Int. J. Multiphase Flow, 27*(10), 1829–1858.

Sørensen, M. (1991), Estimation of some aeolian saltation transport parameters from transport rate profiles, *Acta Mech., 1*, 141–190.

Soulsby, R. (1997), *Dynamics of Marine Sands: A Manual for Practical Applications*, Thomas Telford, London.

Spies, P.-J., I. K. McEwan, and G. R. Butterfield (2000), One-dimensional transitional behaviour in saltation, *Earth Surf. Processes Landforms, 25*, 505–518.

Statham, I. (1976), A scree slope rockfall model, *Earth Surf. Processes, 1*, 43–62.

Stoica, P., and N. Sandgren (2006), Spectral analysis of irregularly-sampled data: Paralleling the regularly-sampled data approaches, *Digital Signal Process., 16*, 712–734.

Stone, M., and A. Murdoch (1989), The effect of particle size, chemistry and mineralogy of river sediments on phosphate adsorption, *Environ. Tech. Lett., 10*, 501–510.

Stout, J. E. (2004), A method for establishing the critical threshold for aeolian sediment transport in the field, *Earth Surf. Processes Landforms, 29*, 1195–1207.

Stout, J. E., and T. M. Zobeck (1997), Intermittent saltation, *Sedimentology, 44*, 959–970.

Strahler, A. N. (1950), Davis' concepts of slope development viewed in the light of recent quantitative investigations, *Ann. Assoc. Am. Geogr., 40*, 209–213.

Strahler, A. N. (1952), Dynamic basis of geomorphology, *Geol. Soc. Am. Bull., 63*(9), 923–938.

Sutherland, A. J. (1987), Static armour layers by selective erosion, in *Sediment Transport in Gravel-Bed Rivers*, edited by C. R. Thorne, J. C. Bathurst, and R. D. Hey, pp. 243–267, John Wiley, Chichester.

Sutter, R. D., R. Verhoeven, and A. Krein (2001), Simulation of sediment transport during flood events: Laboratory work and field experiments, *Hydrol. Sci. J., 46*(4), 599–610.

Swart, D. H., and C. A. Fleming (1980), Longshore water and sediment movement, ASCE Proceedings of the 17th Conference on Coastal Engineering, 1275–1294.

Sweeney, M. R., and A. Mason (2013), Mechanisms of dust emission from Pleistocene loess deposits, Nebraska, USA, *J. Geophys. Res. Earth Surf., 118*, 1460–1471, doi:10.1002/jgrf.20101.

Takahashi, T. (1978), Mechanical characteristics of debris flow, *J. Hydraul. Div., Am. Soc. Civil Eng., 104*, 1153–1169.

Takahashi, T. (1980), Debris flow on prismatic open channel, *J. Hydraul. Div., Am. Soc. Civil Eng., 106*, 381–396.

Takahashi, T. (1991), *Debris Flow*, 1st ed., 165 pp., Balkema, Rotterdam.

Takahashi, T. (2009), A review of Japanese debris flow research, *Int. J. Erosion Control Eng., 2*, 1–14.

Takahashi, T., H. Nakagawa, and S. Kuang (1987), Estimation of debris flow hydrograph on varied slope bed, in *Erosion and Sedimentation in the Pacific Rim (Proceedings of the Corvallis Symposium, August, 1987)*, IAHS Publ., vol. 165, edited by R. L. Beschta et al., pp. 167–177, IAHS, Wallingford.

Tanner, W. F. (1979), k-type coefficients in the littoral drift system, *Mar. Geol., 33*, M15–M19.

Thompson, C., and J. Croke (2008), Channel flow competence and sediment transport in upland streams in southeast Australia, *Earth Surf. Processes Landforms, 33*(3), 329–352.

Toffaleti, F. B. (1968), A procedure for computation of the total river sand discharge and detailed distribution, bed to surface, Tech. Rep. 5, US Army Corps of Engineers, Vicksburg, Miss.

Torri, D. (1987), A theoretical study of soil detachability, *Catena Suppl., 10*, 15–20.

Tunnicliffe, J., A. S. Gottesfeld, and M. Mohamed (2000), High resolution measurement of bedload transport, *Hydrol. Processes, 14*, 2631–2643.

Turowski, J. M., D. Rickenmann, and S. J. Dadson (2010), The partitioning of the total sediment load of a river into suspended load and bedload: A review of empirical data, *Sedimentology, 57*, 1126–1146.

Ungar, J. E., and P. K. Haff (1987), Steady state saltation in air, *Sedimentology, 34*, 289–299.

USDA (1995), USDA-Water Erosion Prediction Project Hillslope Profile and Watershed Model Documentation, NSERL Rep. 10, USDA-ARS-MWA, West Lafayette, In.

van Rijn, L. C. (1984a), Sediment transport. Part I: bed load transport, *J. Hydraul. Eng., 110*, 1431–1456.

van Rijn, L. C. (1984b), Sediment transport. Part II: Suspended load transport, *J. Hydraul. Eng., 110*, 1613–1641.

van Rijn, L. C. (1993), *Principles of Sediment Transport in Rivers, Estuaries and Coastal Seas*, Aqua Publ., Amsterdam.

van Rijn, L. C. (2014), A simple general expression for longshore transport of sand, gravel and shingle, *Coastal Eng., 90*, 23–39.

Vandaele, K., and J. Poesen (1995), Spatial and temporal patterns of soil erosion rates in an agricultural catchment, central Belgium, *Catena, 25*(1–4), 213–226.

Vanoni, V. A. (1975), Sedimentation Engineering. ASCE Manuals and Reports on Engineering Practice, N°. 54, Reston, Va.

Wainwright, J., and J. B. Thornes (1991), Computer and hardware simulations of archæological sediment transport, in *Computer Applications and Quantitative Methods in Archæology 1990*, BAR Int. Ser., vol. 565, edited by K. Lockyear and S. Rahtz, pp. 183–194, BAR Int., Oxford.

Wainwright, J., A. J. Parsons, E. N. Müller, R. E. Brazier, D. M. Powell, and B. Fenti (2008), A transport-distance approach to scaling erosion rates: 1. Background and model development, *Earth Surf. Processes Landforms, 33*, 813–826.

Wainwright, J., A. J. Parsons, E. N. Müller, R. E. Brazier, and D. M. Powell (2010), Standing proud: A response to "Soil-erosion models: Where do we really stand?" by Smith et al, *Earth Surf. Processes Landforms, 35*, 1349–1356.

Walker, I. J. (2005), Physical and logistical considerations using ultrasonic anemometry in aeolian sediment transport research, *Geomorphology, 68*, 57–76.

Warburton, J., and T. R. H. Davies (1998), Use of hydraulic models in management of braided gravel-bed rivers, in *Gravel-Bed Rivers in the Environment*, edited by P. C. Klingeman et al., pp. 513–542, Water Resour. Publ. LLC, Highlands Ranch, Colo.

Weertman, J. (1957), On the sliding of glaciers, *J. Glaciol., 3*, 33–38.

Weertman, J. (1964), The theory of glacier sliding, *J. Glaciol.*, *5*, 287–303.

White, B. R., and H. Mounla (1991), An experimental study of Froude number effect on wind-tunnel saltation, *Acta Mech. Suppl.*, *1*, 145–157.

White, W. R., R. Bettess, and E. Paris (1982), Analytical approach to river regime, *J. Hydraul. Div., ASCE*, *108*(10), 1179–1193.

Whiting, P. J., and J. G. King (2003), Surface particle sizes on armoured gravel streambeds: Effects of supply and hydraulics, *Earth Surf. Processes Landforms*, *28*, 1459–1471.

Whiting, P. J., W. E. Dietrich, L. B. Leopold, T. G. Drake, and R. L. Shreve (1988), Bedload sheets in heterogeneous sediment, *Geology*, *16*, 105–108.

Wiberg, P. L., and J. D. Smith (1985), A theoretical model for saltating grains in water, *J. Geophys. Res.*, *90*, 7341–7354, doi:10.1029/JC090iC04p07341.

Wilcock, P. R. (1998), Two-fraction model of initial sediment motion in gravel-bed rivers, *Science*, *280*, 410–412.

Wilcock, P. R. (2001), Toward a practical method for estimating sediment transport rates in gravel-bed rivers, *Earth Surf. Processes Landforms*, *26*, 1395–1408.

Wilcock, P. R., and J. C. Crowe (2003), Surface-based transport model for mixed-size sediment, *J. Hydraul. Eng.*, *129*, 120–128.

Wilcock, P. R., and B. T. DeTemple (2005), Persistence of armor layers in gravel-bed streams, *Geophys. Res. Lett.*, *32*, L08402, doi:10.1029/2004GL021772.

Wilcock, P. R., and B. W. McArdell (1993), Surface-based fractional transport rates: Mobilization thresholds and partial transport of a sand-gravel sediment, *Water Resour. Res.*, *29*, 1297–1312, doi:10.1029/92WR02748.

Wilcock, P. R., and J. B. Southard (1989), Bed load transport of mixed size sediment: Fractional transport rates, bed forms, and the development of a coarse bed surface layer, *Water Resour. Res.*, *25*, 1629–1641, doi:10.1029/WR025i007p01629.

Wilcock, P. R., S. T. Kenworthy, and J. C. Crowe (2001), Experimental study of the transport of mixed sand and gravel, *Water Resour. Res.*, *37*, 3349–3358, doi:10.1029/2001WR000683.

Williams, G. (1964), Some aspects of the aeolian saltation load, *Sedimentology*, *29*, 409–417.

Williams, G. P. (1989), Sediment concentration versus water discharge during single hydrologic events in rivers, *J. Hydrol.*, *111*, 89–106.

Wilson, K. C. (1966), Bedload transport at high shear stress, *J. Hydraul. Div., Am. Soc. Civil Eng.*, *92*, 49–59.

Wilson, K. C. (1989), Mobile-bed friction at high shear stress, *J. Hydraul. Eng.*, *115*, 825–830.

Woolhiser, D. A., R. E. Smith, and D. C. Goodrich (1990), KINEROS: A kinematic runoff and erosion model—Documentation and user manual, U.S. Department of Agriculture, Agricultural Research Service, ARS-77, 130 pp.

Wright, L. D., J. Chappell, B. G. Thom, M. P. Bradshaw, and P. J. Cowell (1979), Morphodynamics of reflective and dissipative beach and inshore systems, southeastern Australia, *Mar. Geol.*, *32*, 105–140.

Xia, J., Z. Wang, Y. Wang, and X. Yu (2013), Comparison of morphodynamic models for the lower Yellow River, *J. Am. Water Resour. Assoc.*, *49*, 114–131, doi:10.1111/jawr.12002.

Yalin, M. S. (1972), *Mechanics of Sediment Transport*, Pergamon Press, Oxford.

Yalin, Y. S. (1963), An expression for bed-load transportation, *J. Hydraulic Div. ASCE*, *89*, 221–250.

Yang, C. T. (1972), Unit stream power and sediment transport, *J. Hydraul. Div., ASCE*, *98*(HY10), 1805–1826.

Yang, C. T. (1973), Incipient motion and sediment transport, *J. Hydraul. Div., ASCE*, *99*(HY7), 919–934.

Yang, C. T. (1984), Unit stream power equation for gravel, *J. Hydraul. Div., ASCE*, *110*(12), 1783–1797.

Yang, S. (2005), Formula for sediment transport in rivers, estuaries, and coastal waters, *J. Hydraul. Eng.*, *131*(11), 968–979.

Zhang, G.-H., Y.-M. Liu, Y.-F. Han, and X. C. Zhang (2009), Sediment transport and soil detachment on steep slopes: I. Transport capacity estimation, *Soil Sci. Soc. Am. J.*, *73*(4), 1291–1297.

Zhang, G.-H., L.-L. Wang, K.-M. Tang, R.-T. Luo, and X. C. Zhang (2011), Effects of sediment size on transport capacity of overland flow on steep slopes, *Hydrol. Sci. J.*, *55*, 1289–1299.

Zingg, A. W. (1940), Degree and length of land slope as it affects soil loss in run-off, *Agric. Eng.*, *21*, 59–64.

Zingg, A. W. (1953), Wind tunnel studies of the movement of sedimentary material, in *Proceedings of the 5th Hydraulics Conference Bulletin Institute of Hydraulics*, pp. 111–135, Univ. of Iowa Studies in Eng., Iowa City.