Bed Material Transport and the Morphology of Alluvial River Channels

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Key Words

alluvial sedimentation, channel geometry, fluvial bedforms, fluvial geomorphology, sediment transport

Abstract

The morphology of an alluvial river channel is the consequence of sediment transport and sedimentation in the river. Morphological style is determined chiefly by the caliber and quantity of sediment delivered to the channel, although modulated by channel scale. Yet the relations between sediment transport and river morphology have received only limited, qualitative attention. In this review, the problem is studied by defining sediment transport regimes on the basis of the Shields number, a nondimensional measure of the capacity of the channel to move sediment of a given caliber. The problem is also approached from an inverse perspective by which the quantity and character of sediment deposits are used to infer details about the variation of sediment transport and sedimentation along a channel. Coupling the two approaches establishes a basis to gain new insights into the origins of alluvial channel morphology.

INTRODUCTION

Bed material: material that forms the bed and lower banks of the river and chiefly determines the morphology of the channel

Wash material: material that, once entrained, is transported for a long distance in suspension

Alluvial river channels are formed in the sediments that they have transported and deposited. Accordingly, the channel is self-formed. Channel geometry and morphology are the direct consequence of the sediment transport process, yet this essential connection has received relatively little emphasis in textbooks (but see Allen 1985, Bridge 2003). The purpose of this review is to explore connections between fluvial sediment transport and river channel form.

Clastic sediments in transport customarily are classified on the basis of the mechanism by which they are moved, the principal categories in rivers being suspension and bed load. Suspended sediment is supported in the water column by upwardly directed turbulent water motions. Such material may travel a long way before being deposited. Bed load (traction transport) progresses by rolling, sliding, or bouncing over the river bottom, its weight remaining principally supported by the bed. Such material is apt to travel only a short distance in one movement. Saltation is a third category of motion in which particles are launched into the water column but then return relatively quickly to the bed following a ballistic trajectory. It is much less important in water than in eolian transport, but is in practice difficult to separate from intermittent suspension.

For considering fluvial sedimentation and river channel morphology, sediments are more appropriately divided into bed material and wash material (see Sidebar for expanded definitions). The former is relatively coarse material that makes up the bed and lower banks of the river channel and is of major importance in determining river channel morphology. The latter is fine material that, once entrained, travels out of the reach. Wash material is not normally found in significant quantity in the bed and lower banks of the channel, although it may occur interstitially within bed material deposits. It is deposited in slack water on bar tops and overbank during floods, and therefore may be an important constituent of the upper banks.

EXPANDED DEFINITIONS OF BED MATERIAL AND WASH MATERIAL

Bed material: material that forms the bed and lower banks of the river and chiefly determines the morphology of the channel. In alluvial channels, it corresponds with the coarser part of the sediment load transported by the river, and it may move either as bedload or as intermittently suspended load.
Wash material: material that, once entrained, is transported for a long distance in suspension. This material is found only in minor quantities (the result of interstitial trapping) in the bed of the river, but may form a significant fraction of upper bank and floodplain deposits as the result of deposition in quiet water overbank during floods. Sediment classified as wash material in one reach of a river may become bed material in another reach with lower sediment transporting power.

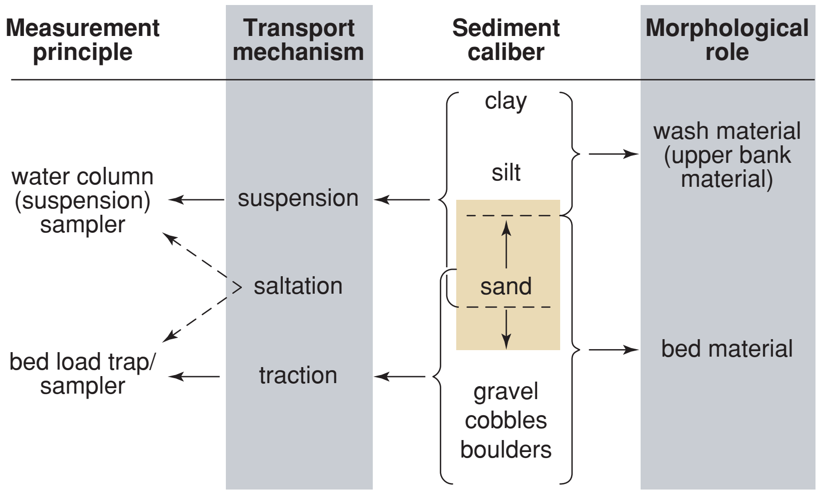

Figure 1
Categorization of fluvial sediments according to transport process, measurement principle, and morphological/sedimentary association.

Alluvial channel: a channel formed by a river in its own sediments, which it has transported and deposited, and is capable of remobilizing

Competence: the ability of a stream flow to mobilize sediment of a given size, quantified by the Shields number, $\tau^* = \rho g d S / g(\rho s - \rho) D$

Bed material is often conflated with bed load, and wash material certainly moves in suspension, but the two classifications are not congruent. Medium and coarse sands, in particular, constitute bed material that may move into suspension in strong currents. Furthermore, sand is common in river systems. Fluvial sediment transport normally is measured according to operational principles that essentially correspond with the definitions of transport process (**Figure 1**). Perhaps this is a reason why relatively little emphasis has been given to the connections between transport and alluvial morphology; the available data of sediment transport do not conveniently lend themselves to the analysis of fluvial sedimentation.

Nevertheless, there certainly have been attempts to understand the relations between sediment transport and alluvial channel morphology. A general association between fine sediments moving in suspension and meandered rivers on low gradients, and a contrasting one between coarse sediments moving in traction and rivers of low sinuosity on relatively high gradients, has been recognized for a long time. In 1963, S.A. Schumm published a table classifying alluvial rivers into three categories on the basis of the dominant mode of sediment transport and giving some characteristics of the channels associated with each. Schumm's classification recognized two end-member channel types associated with suspended load and with bed load dominance, as above, and a mixed load category between (see **Table 1**). Emphasis on the transport process assigns appropriate prominence to the competence (see Sidebar for expanded definition) of the river to move sediment, hence to the power of the river.

Somewhat more formally, the American hydraulic engineer E.W. Lane, in the 1950s, represented the equilibrium or graded condition for an alluvial river channel (the condition in which it passes the imposed water and sediment fluxes without net change in form) by the simple qualitative statement $QS \sim Q_s D$, in which Q is discharge, S is channel gradient, Q_s is sediment flux (in fact, the most appropriate definition is bed material flux), and D is sediment caliber (Lane 1955). The relation states that, for given flow energy, so much sediment of some specified

Table 1 Elementary classification of alluvial river channels and riverine landscapes[a]

Type/characteristic Shields number[b]	Sediment type	Sediment transport regime	Channel morphology	Channel stability
Jammed channel 0.04+	Cobble- or boulder-gravel	Bed load dominated; low total transport, but subject to debris flow	Step-pools or boulder cascades; width typically a low multiple of largest boulder size; $S > 3°$	Stable for long periods with throughput of bed load finer than structure-forming clasts; subject to catastrophic destabilization in debris flows
Threshold channel 0.04+	Cobble-gravel	***Bed load dominated***; low total transport in partial transport regime; bed load may actually be less than 10% of total load	Cobble-gravel channel bed; single thread or wandering; highly structured bed; relatively steep; low sinuosity; $w/d > 20$, except in headwater boulder channels	Relatively stable for extended periods, but subject to major floods causing lateral channel instability and avulsion; may exhibit serially reoccupied secondary channels
Threshold channel up to 0.15	Sandy-gravel to cobble-gravel	Bed load dominated, but possibly high suspension load; partial transport to full mobility; bed load typically 1%–10% of total load	Gravel to sandy-gravel; single thread to braided; limited, local bed structure; complex bar development by lateral accretion; moderately steep; low sinuosity; w/d very high (>40)	Subject to avulsion and frequent channel shifting; braid-form channels may be highly unstable, both laterally and vertically; single-thread channels subject to chute cutoffs at bends; deep scour possible at sharp bends
Transitional channel 0.15–1.0	Sand to fine-gravel	***Mixed load***; high proportion moves in suspension; full mobility with sandy bedforms	Mainly single-thread, irregularly sinuous to meandered; lateral/point bar development by lateral and vertical accretion; levees present; moderate gradient; sinuosity <2; $w/d < 40$	Single-thread channels, irregular lateral instability or progressive meanders; braided channels laterally unstable; degrading channels exhibit both scour and channel widening
Labile channel >1.0	Sandy channel bed, fine-sand to silt banks	***Suspension dominated*** with sandy bedforms, but possibly significant bedload moving in the bedforms	Single thread, meandered with point bar development; significant levees; low gradient; sinuosity >1.5; $w/d < 20$; serpentine meanders with cutoffs	Single-thread, highly sinuous channel; loop progression and extension with cutoffs; anastomosis possible, islands are defended by vegetation; vertical accretion in the floodplain; vertical degradation in channel

(Continued)

Table 1 *(Continued)*

Type/characteristic Shields number[b]	Sediment type	Sediment transport regime	Channel morphology	Channel stability
Labile channel up to 10	Silt to sandy channel bed, silty to clay-silt banks	Suspension dominated; minor bedform development; minor bed load	Single-thread or anastomosed channels; prominent levees; very low gradient; sinuosity > 1.5; w/d < 15 in individual channels	Single-thread or anastomosed channels; common in deltas and inland basins; extensive wetlands and floodplain lakes; vertical accretion in floodplain; slow or no lateral movement of individual channels

[a]Bold, italic entries were included in Schumm's original classification. Some of the parameters in the descriptions have been changed, in light of further experience, from Schumm's original values.
[b]Values apply to channel-forming (i.e., flood) flows.

size can be transported. This is a qualitative statement of the principal governing conditions of alluvial channel form. Discharge chiefly determines the scale of the channel and gradient determines the rate of energy expenditure, whereas, for the given scale and gradient, the character of alluvial morphology is chiefly determined by the caliber and quantity of sediment delivered to the channel. The balance of the governing conditions also determines the stability of the channel—that is, the propensity for aggradation or degradation and the style and rate of lateral movement. Morphology and stability have been represented in a diagram that approximately relates channel morphology to the sedimentary governing conditions (Mollard 1973, Schumm 1985). **Figure 2** presents an evolved version of the diagram. The associations presented in the diagram have not, however, been placed on a physically firm foundation.

We define our problem as determining the expected morphology of an alluvial river channel, given some particular governing conditions. We seek to explain the

EXPANDED DEFINITION OF COMPETENCE

Competence: the ability of a stream flow to mobilize sediment of a given size, quantified by the Shields number, $\tau^* = \rho g d S / g(\rho_s - \rho)D$, a dimension-free measure of the shear stress exerted by the flow on the bed, in which ρ_s and ρ are sediment and fluid densities, respectively; g is the acceleration of gravity; d is flow depth; and D is the grain size to be moved. For a closely packed population of grains of similar size, the critical value for entrainment is $\tau_c^* \approx 0.06$, but for usual mixtures of sediments on stream beds, it has been found that $\tau_c^* \approx 0.045$ when D is D_{50} of the surface material, and that all sizes move off within a narrow range of τ^*. This occurs because larger grains shelter smaller ones, so that once the larger grains begin to move, all grains may be entrained.

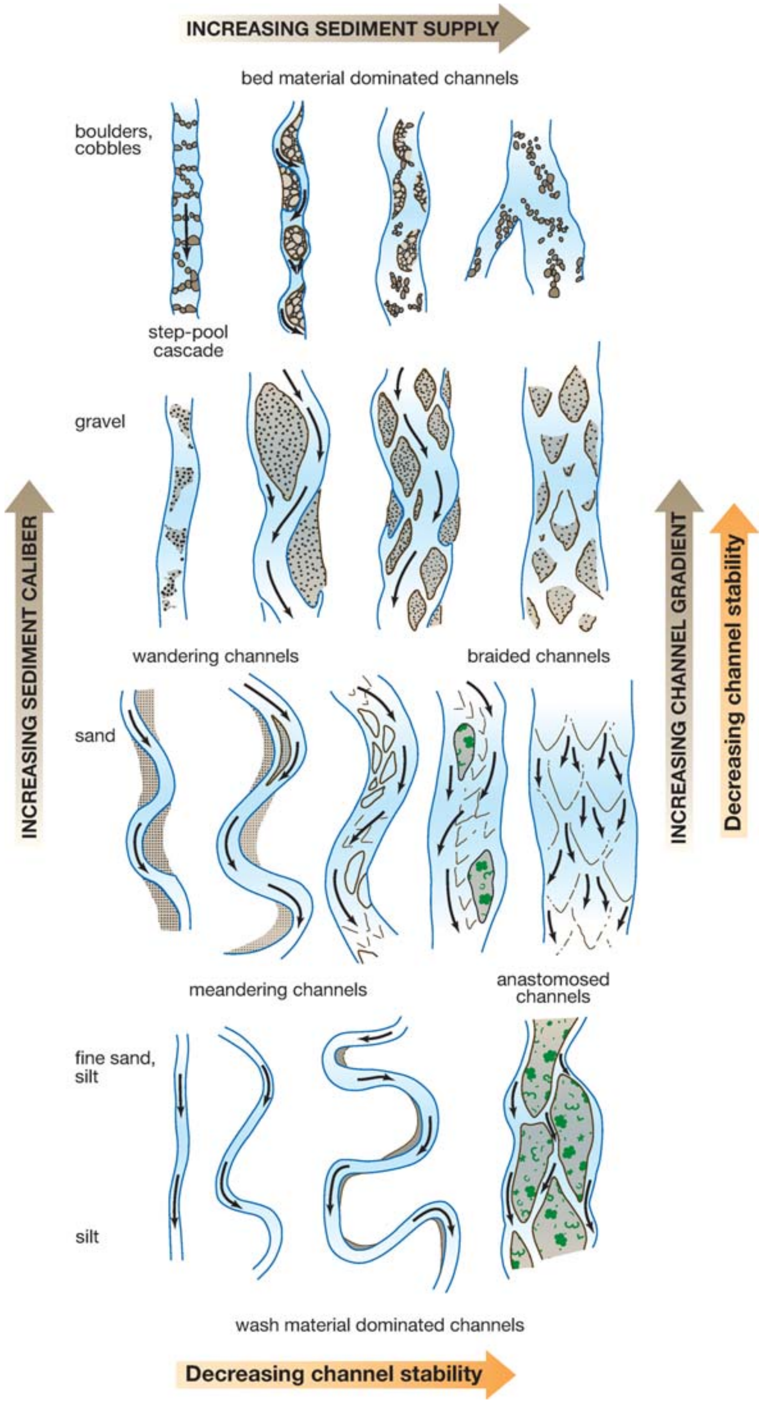

Figure 2

Diagram showing the association of alluvial river channel form and the principal governing factors (modified after Church 1992, based on the concept of Mollard 1973 and Schumm 1985). Classically named channel types are located at appropriate positions within the diagram. Shading is intended to reflect sediment character.

patterns implicit in **Figure 2**. The problem is approached in two ways. In a forward approach to the problem, we use known physics of sediment transport to deduce some conditions of fluvial sedimentation. In an inverse examination, we use observed properties of stream channels and fluvial sediments to make inferences about the sediment transport process. The forward approach can be applied relatively rigorously, but leads—in the present state of knowledge—only to rather general results. The inverse approach, which is attractive because river morphology and river sediments are much more easily observed than sediment transport, can yield quite detailed results on along-channel variations in transport, hence morphology, but is not yet so physically rigorous.

THE FORWARD PROBLEM

Basic Relations

Recasting Lane's relation as $QS \sim Q_b D/d$, in which Q_b is bed material load and d is flow depth, we obtain a relation that recognizes that it is the relative scale of flow and sediment that effectively controls stream competence and is also rational. We find, upon rearrangement and insertion of appropriate constants,

$$Q_b/Q = f[\rho g d S/g(\rho_s - \rho)D], \tag{1}$$

which is Shields's (1936) function, g being the acceleration of gravity, and ρ_s and ρ are the density of sediment and water, respectively. The term in brackets is conventionally notated by τ^* and is referred to as the Shields number. It is a dimensionless expression, scaled to grain size, of the shear stress imposed by the flow. The term on the left is the average concentration of bed material sediment in transport, so Equation 1 is a sediment transport function. It has been found by experiment (Meyer-Peter & Müller 1948, Wilcock & Southard 1988, review in Komar 1996) that, in the limit when $Q_b/Q \to 0$, for widely graded, unconstrained sediment mixtures, $\tau^* \equiv \tau_c^* \to 0.045$. This condition defines the competence of the stream with respect to grain size.

Dade & Friend (1998) have explored the characteristic behavior of τ^* for various grades of sediment in transport. Beginning from the Rouse equation (for the distribution of suspended sediment in a shear flow), and assuming that the required reference sediment concentration at the base of the profile is the concentration in the bed load, they obtained characteristic relations between $w_s/u*$, the ratio of settling velocity to shear velocity of the flow, and the fraction of the load carried as bed load (**Figure 3**). The ratio varies with relative depth, $\zeta = d/z_b$, in which d is flow depth and z_b is the thickness of the bed load layer. For gravels, ζ is approximately the inverse of relative roughness (*D/d* in Lane's relation) because $z_b \approx D$. They defined limits of $w_s/u* \leq 0.3$ for dominantly suspended transport and $w_s/u^* \geq 3$ for dominant bed load. Domination by suspension is considered to occur when bed load constitutes less than 10% of the load. These are arbitrary figures, however, which must be compared with experience (see **Table 1**). The comparison is difficult because the division between suspended and bed load customarily is given in terms of the total sediment load, so the sometimes abundant wash load substantially inflates the suspended fraction. In

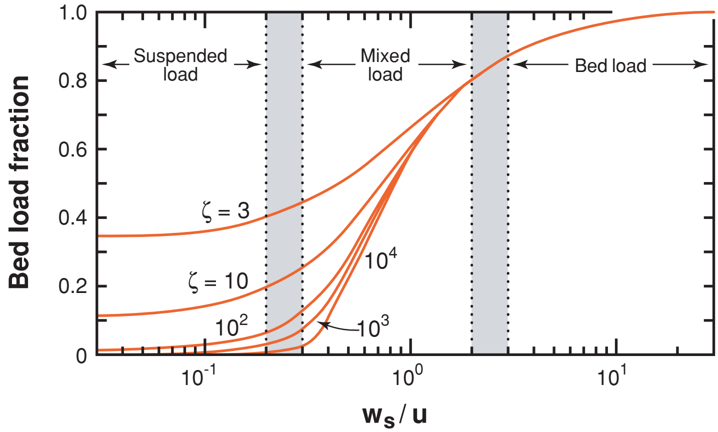

Figure 3

Fraction of the total sediment load carried as bed load portrayed as a function of settling velocity ratio, $w_s/u*$, and relative depth, $\zeta = d/z_b$ (extended after figure 1 in Dade & Friend 1998). Limits for transport regime types are as quoted in Dade & Friend and revised in Dade (2000).

comparison, it appears as if Dade and Friend's fractions, which—strictly—should be applied to bed material, inflate the bed load fraction.

Dade and Friend went on to apply a conceptual sediment transport equation similar to that of Bagnold (1966) to infer the fluid stress regime that leads to dominance by one component of the load or the other. Expressed in terms of the Shields number, so that sediment characteristics are incorporated, it appears that $\tau* \approx O[1]$ to drive a suspension-dominated system in equilibrium (that is, to maintain transport with no net erosion or deposition), and $\tau* \rightarrow \tau*_c$ for a bed load–dominated system. For mixed load systems, $3\tau*_c < \tau* < O[1.0]$, approximately. Data displayed in a Shields-type diagram (**Figure 4**) demonstrate these regimes.

The Shields number can be rearranged to yield, when the constants are replaced,

$$S = 1.65\tau* D/d. \tag{2}$$

The association of this functional relation with characteristic sediment transport regimes is illustrated in **Figure 5**. Evidently, gradient, scale, and sediment properties determine the transport regime for a particular stream, as predicted from the governing conditions. All of the factors except scale can be adjusted by the river, but not quickly. It can be inferred from this relation, as well, that sediment transport regimes must change systematically through a normal drainage basin as gradient and relative roughness decline.

Ashworth & Ferguson (1989) and Warburton (1992), on the basis of observations in gravel-bed rivers, identified three phases of bed material transport: an overpassing phase, in which material—almost always sand—advected from upstream is moved over a static local bed; a size-selective phase, in which only some of the clasts on the local bed are entrained (see Komar 1996); and a fully mobile phase, in which all material on the local bed participates in the transport (see also Jackson & Beschta 1982). Wilcock & McArdell (1993) defined a regime of partial transport, in which only some of the grains on the bed are mobile at any given time, even though all sizes may eventually participate in the motion. It is evident from field studies (Church & Hassan 2002) that both partial and size-selective transport occur near the threshold

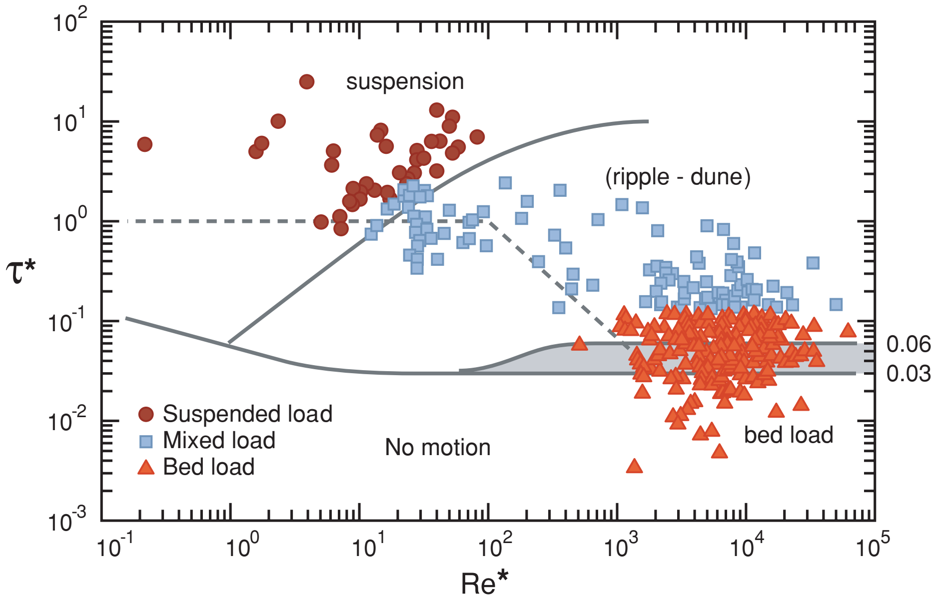

Figure 4

The Shields diagram with data for bed load (threshold channels), mixed load (transitional channels), and suspended load (labile channels) dominance superimposed (after Dade & Friend 1998). The abscissa is $Re* = u*D/\nu$, the grain Reynolds number, in which ν is the kinematic viscosity of water. It can be interpreted as a scaled grain size. The threshold of motion at high $Re*$ is represented for Shields' classical limit (0.06) and is a frequently quoted lower value for motion (0.03); the usually accepted value for sediment mixtures falls midway between. The boundary for sediment suspension is not precise. The data are for many of the same rivers as presented by Dade & Friend (1998, figure 2) but are calculated for mean annual flood or bankfull flow because the dominant mode of sediment transport is set by high flows (Dade & Friend presented data for mean annual flow).

for bed material motion. Full mobility, on the other hand, is characteristic of mixed and suspended load streams, and it is usefully divided into tractive and suspended phases.

These bed material transport phases may be collated with the Shields number ranges of Dade and Friend to create a classification of alluvial river channels according to the dominant mode of sediment transport. This substantially elaborates Schumm's (1963) classification of alluvial channel types (**Table 1**) and goes some considerable way toward providing the foundation for the range of river channel morphologies displayed in **Figure 2**. The Dade and Friend analysis is, of course, highly reductionist, and the full range of river behaviors remains more varied, as the elaborations of **Table 1** and **Figure 2** show.

We can, then, define a class of channels (**Figures 4** and **5**; **Table 1**) in which bed material transport is restricted to conditions near threshold ($O[0.01 \leq \tau* \leq 0.1]$). Such channels correspond with Dade & Friend's bed load type and with the partially mobile

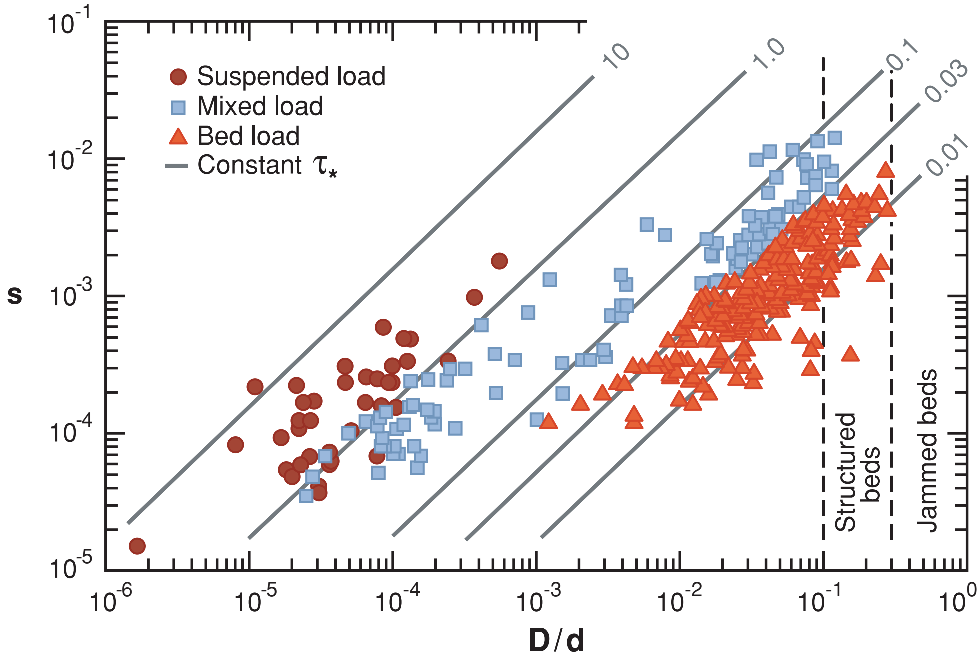

Figure 5

River data displayed in a graph of channel gradient (*S*) versus relative roughness (*D*/*d*). Lines of constant $\tau*$ are straight lines in this diagram. Data of bed load (threshold) and closely associated mixed (transitional) channels plot lower in this diagram, in relation to $\tau*$, than would be expected (compare with **Figure 4**) because the actual grain sizes in transport in these channels are customarily smaller than the grain sizes exposed on the bed surface. The estimate of relative roughness is based on the latter because transport data are available for very few of the channels. The discrepancy is a factor of approximately 3.

condition of Wilcock & McArdell. They may be classified as threshold channels, and they occur in gravels and coarser materials. Another well-defined class is made up of channels in which bed material moves into suspension, corresponding with Dade & Friend's suspended load type ($\tau* > O[1]$); they certainly experience full mobility and may be classified as labile channels because the bed is apt to be deformed continuously by the sediment transport process. Such channels occur in sands and finer sediments. Between these two clearly distinguished classes is a group of transitional channels ($O[0.1] \leq \tau* \leq O[1]$) commonly occurring in fine gravels or sandy gravels, or in sand-bed channels on low gradients, in which full bed material mobility occurs but a significant portion of the sediment load remains in traction. (See Sidebar for expanded definition of alluvial channel types.)

Threshold channel: river channel in which the limit of competence for bed material transport is characteristically exceeded by only a modest amount

Labile channel: river channel in which the bed sediments are relatively easily and frequently entrained by the flow

The three transport regimes described above are clearly equivalent to Schumm's original three categories, but focus on the range of Shields numbers permits a finer division of channel regime types (**Table 1**). The dominant sediment transport mode controls the nature of sediment accretion and, consequently, the major features of

EXPANDED DEFINITIONS OF ALLUVIAL CHANNEL TYPES

Labile channel: river channel in which the bed sediments are relatively easily and frequently entrained by the flow, typically channels with sand beds ($\tau* = O[1]$ at high flows). Morphological changes may be relatively rapid, but, in labile channels, lateral instability is often strongly constrained by strong banks reinforced by vegetation.

Threshold channel: river channel in which the limit of competence for bed material transport is characteristically exceeded by only a modest amount, so that the transport of bed material typically occurs only at low intensity ($\tau* = O[0.1]$ at high flows). Partial transport (only some of the grains on the bed are in motion at any time) and size-selective transport (larger grains may not move at all) are characteristic of threshold channels, which have bed material composed of gravel or cobbles. Morphological change is slow.

Transitional channel: channel with characteristics intermediate between those of threshold channels and labile channels, typically sandy channels with low energy or gravel-bed channels. Sediment transporting events that mobilize most of the bed material occur moderately frequently, along with associated morphological changes ($O[0.1] < \tau* < O[1]$ at high flows).

channel morphology. Bed load necessarily is deposited as within-channel accumulations around which the stream must flow, giving rise to relatively wide, shallow channel zones and a lateral style of instability that characterizes both threshold and transitional channels. In contrast, suspended sediments are deposited from the water column onto accumulating sediment surfaces that build vertically. Finer, more cohesive sediments lend strength to stream banks and fine sediments encourage rapid establishment of vegetation, therefore labile channels, characteristically, are relatively deeper and narrower, with high bar-scale topography. A wide intermediate class of channels experiences both these modes of sedimentation.

Transitional channel: intermediate between threshold channels and labile channels. Sediment transporting events that mobilize most of the bed material occur moderately frequently

Threshold Channels

It appears as if the driving force for sediment transport in bed load–dominated channels never rises far above the threshold for motion (see Parker & Klingeman 1982, Wilcock & McArdell 1997). From Equation 2, we notice that for any value of relative roughness (that is, relative scale of sediment and flow), there is a limiting slope (S_c) for particle stability set by the threshold Shields number. For $D/d \approx 1$ and $\tau*_c = 0.045$, $S_c \approx 0.07$ (4°). Sediment with relative roughness 1 should not be found in channels with higher gradient. But it certainly is. Channels with gradients as high as 0.2 (11°) retain clastic accumulations. For $D/d < 1$, calculated S_c is even lower. On higher gradients, sediment accumulations take a special form: The largest clasts are locked in jammed structures that form steps and intervening pools. The steps not only immobilize the large sediments but modify energy dissipation—concentrating it in falls—so that, in normally recurring flood flows, only relatively small sediments are moved downstream (Zimmermann & Church 2001). Grant et al. (1990) and Montgomery & Buffington (1997) have classified a range of distinctive morphologies on successively steeper gradients, the most striking of which are step-pools (Chin 1989, Curran & Wilcock 2005) and boulder cascades. Channels exhibiting these morphologies are small in scale and occupy steep gradients in drainage basin headwaters. Their occurrence, of course, depends on the availability of appropriately large key materials (which may be wood, as well as cobbles or boulders). Otherwise, scoured, rock-bound channels occur.

Structural reinforcement occurs on lower gradients in gravel stream beds, with $0.3 < D/d < 1$, where $\tau*$ rises to order 0.1. In these wake-dominated flows, stone clusters (Brayshaw 1984), stone lines (Laronne & Carson 1976, Martini 1977), and stone nets (Church et al. 1998) composed of the largest sediments form on the streambed, reducing sediment mobility by orders of magnitude. Equation 2 indicates that such structures should be found on gradients as low as approximately 1° (i.e., for $D/d \approx 0.3$ and $\tau^* = 0.045$, $S_c \approx 0.02$, or 1°). More generally, gravelly fluvial sediments pervasively exhibit imbrication (Johansson 1976). The parameterization of these conditions to study their effect on sediment transport has been pursued mainly in terms of the observed coarsening of the surface layer of sediments (Parker & Klingeman 1982, Dietrich et al. 1989) and consideration of pivoting angles to remove sediment from imbricated positions (Komar & Li 1988). Surface coarsening yields armor ratios within the range $2 \leq D_{50s}/D_{50b} \leq 4$, in which the grain sizes represent the median size of surface and subsurface (bulk) sediments, respectively. But the mobility of the bed surface depends not just on the surface grain size but also on the surface packing arrangement; that is, on bed structure (see Sidebar). An upward adjustment of τ_c*, which recognizes the Shields number as a bed state parameter rather than simply as a grain mobility index, represents the fundamental approach to this problem. Values of τ_c* as high as 0.075 have been observed experimentally (Buffington & Montgomery 1997, Church et al. 1998) and values greater than 0.1 have been observed in the field (Mueller et al. 2005).

The development of structured streambeds depends on the maintenance of low rates of sediment transport. The largest clasts rarely or never move, so they become the keystones for structure development as other clasts of similar size fetch up against them. Such channels normally transport far less sediment than the theoretical maximum transport—that is, the transport over an unstructured bed composed of the same material—for the bulk sediments present in the bed (which must more nearly represent the long-term average size distribution of sediments passing through the channel than does the surface). The mobile sediment typically is also finer than the material exposed on the bed surface. However, it is true that, near the threshold for sediment motion, the rate of change of sediment transport with increasing flow (and shear force) is very rapid. Hence, such streams are apt to be changed dramatically by high transport—in particular, transport extending to full mobility—in exceptional floods. This point is well illustrated by ephemeral desert channels subject to episodic major floods, which experience high transport and display little or no modification of the surface sediment (Reid & Laronne 1995).

BED STRUCTURE

The grain-on-grain arrangement of an alluvial channel bed, which may influence the propensity for individual grains to be entrained by the flow, is collectively termed bed structure. Bed structure includes several distinct aspects of grain arrangement:
(*a*) Grain packing: the arrangement of grain positions with respect to each other, which is influenced by size and shape of individual grains. Tight packing reduces the propensity for individual grains to be mobilized.
(*b*) Grain imbrication: the "shingling" effect of one grain on another as the result of deposition in a directed flow. Grains typically lie with an upstream dip and the weight of the partly covering, upstream grain discourages mobilization of the downstream grain.
(*c*) Grain clusters: the accumulation of relatively large grains into compact groups of two or more grains on the streambed. This feature is common in threshold channels where immobile large grains form effective blocks against which other grains come to rest.
(*d*) Grain nets: the extension of clusters into irregular lines and cell-like arrangements, sometimes seen in threshold channels.
Grain clusters and grain nets reduce bed material transport by discouraging mobilization of individual grains from the structure and by carrying a large fraction of the total fluid force imposed on the bed, so that smaller grains are sheltered below them.

The channel-scale morphological consequence of individual clast movements is subdued topography with shallow pools (except at forced bends), long rapid sections, and development of generally low alternating bars. Gravel sheets (Whiting et al. 1988, Bennett & Bridge 1995) occur on bar surfaces, the relict of large flows, while bar tops and protected lee sides may have accumulations of finer sediment, including sand (Bluck 1982, Laronne et al. 2001). The three-dimensional sedimentology of such channels remains essentially uninvestigated but is almost certainly restricted to poorly defined, subhorizontal bedding with wide grain size distributions, usually framework supported. Such channels are characteristically found in upland valleys where steep gradients and the influx of large clasts into channels of modest scale establish the conditions for threshold transport regimes and channel morphologies. Where sediment supply is abundant, coarse, braided channel deposits typically occur (e.g., Lunt & Bridge 2004).

Transitional Channels

This class of channels has not been defined closely, so the first question is, What types of channels are included? Schumm identified them only as channels with a high proportion of suspended load and sandy bedforms. Specification of Equation 2 for

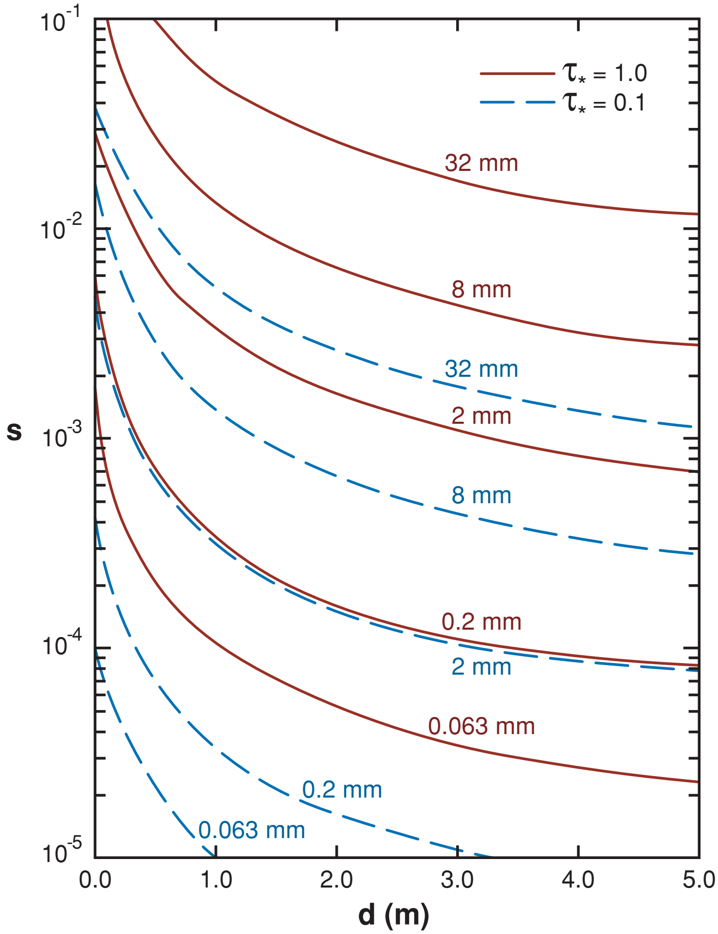

Figure 6

Bed material grain size for various depth-slope combinations in transitional channels. Grain size contours are common textural divisions, except 0.2 mm, which is included because it represents the approximate lower limit for unflocculated material to be found in an open channel bed.

$0.1 \leq \tau* \leq 1.0$ and gradients $S \leq 0.02$, or about 1° (**Figure 6**), shows that channels may vary from cobble-gravel to silt floored, depending on the scale of the channel (here indexed by d). At $\tau* = 0.1$, most of the plausible range is occupied by gravel-bed channels, except for small channels or ones on very low gradients, whereas at $\tau* =$ 1.0, sand-bed channels are included. This outcome is conditioned by the occurrence of grain size as a scale in the determination of $\tau*$.

The morphology and deformation style of transitional channels remain strongly influenced by the bed load, but deposition from suspension also contributes to the production of channel morphology. Wilcock & McArdell (1997) estimated that full mobility occurs when $\tau \approx 4\tau_c$, that is, when $0.12 \leq \tau* \leq 0.18$ (see **Figure 4**). Bed material moves as gravel or sand sheets and sands go into suspension, but, especially in channels with flood conditions nearer $\tau* \approx 0.2$, full mobility is realized relatively rarely—perhaps on the order of 0.1% of time. In sandy channels ($\tau* = O[1]$), full mobility is apt to occur more frequently and, because the sand moves as a grains-deep

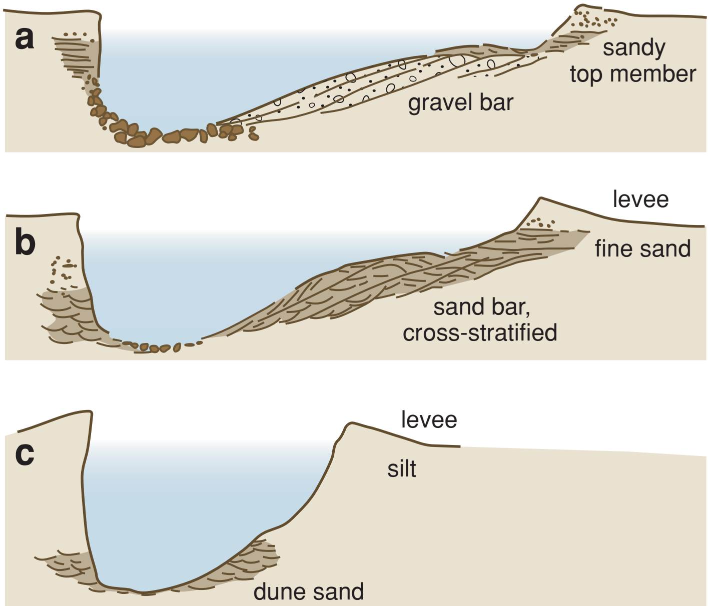

Figure 7

Cartoons for sedimentary style in transitional rivers. See text for discussion.

layer, instability on the water/bed layer interface spontaneously generates bedforms. The bed typically is rippled or dune-covered. Channels of this type are found on moderate gradients in trunk valleys where the supply of mobilizable sediment is relatively abundant.

Larger values of $\tau*$ are systematically associated with smaller values of relative roughness. Once D/d declines below approximately 0.1, substantial vertical accumulations of sediment may occur, and so vertically stacked sediments appear with characteristic upward fining. Gravel is deposited in the channel bed as tabular or inclined bar forms that develop from stalled sheets. Finer gravels form accumulations with avalanche fronts much as do coarse sands, yielding cross-set bedding. The river must find its way around such accumulations so that a lateral style of instability results, bar accumulation being matched by bank erosion in similarly noncohesive materials deposited earlier (**Figure 7*a***). Sands are swept onto bar tops or deposited there from suspension, adding depth to the sediment pile. In sandy environments (see Todd 1996), the entire sediment pile consists of upward-fining sands (**Figure 7*b***). The bed load–driven lateral sedimentation style means that flanking floodplains develop by lateral accretion, with a more-or-less significant topstratum of suspension-derived sands deposited during overbank flood (see floodplain class B in Nanson & Croke 1992). In small channels with relatively strong, silty banks, lateral scour and vertical fill may represent the main erosion/deposition mechanisms associated with bar growth and trimming and with sand wave propagation (**Figure 7*c***). (Bank strength depends on sediment properties, not flow scale, but in smaller channels, the fluid erosional forces that may be brought to bear on the banks are smaller so that the banks are less easily eroded.) Sedimentary style in transitional channels has been extensively pursued (see Bridge 2003, pp. 214–38, for a recent review).

Planform styles of transitional and labile channels have been extensively investigated empirically, but the main effort has been expended on attempts to find a discriminating criterion between meandered (single-thread) and braided (multi-thread) channels [Bridge (2003) gives a review, and recent work has been published by Millar (2000) and Lewin & Brewer (2001)]. Planform must depend on the interaction of the governing conditions flow (Q) and imposed bed material load (Q_b) (or sediment concentration), available valley gradient (S_v), and bank material strength. If the imposed sediment concentration is greater than can be moved on the available gradient, then the channel must deposit part of its load and it will generally aggrade and braid. Bed load–transporting channels may also braid at grade when banks are noncohesive, as has been shown in laboratory experiments (e.g., Ashmore 1982, 1991). If the imposed load is smaller than can be moved on the available gradient, then the river will entrain sediment and degrade. The preferred mode of transient degradation is for the channel to become more sinuous until the channel gradient is reduced to the requisite value (see Bettess & White 1983, Eaton et al. 2004) unless bank strength prevents it, but even in this circumstance, low-order braiding may occur if the channel is transporting material only in traction, so that there is no upper-bank construction on the inside of developing bends. (The term degradation is used here in the most general sense to indicate net evacuation of sediment from the channel, not just in the restricted sense of vertical incision.)

Some important distinctions in channel deformation style appear to be systematically associated with sediment caliber and sedimentation. Alternate bars progress along nearly straight channels (Whiting & Dietrich 1993), and, similarly, meanders in gravel or coarse sand, with relatively noncohesive banks, progress along the channel. In comparison, meanders in channels with more cohesive banks tend to become anchored and to develop by loop extension and rotation, yielding more complex patterns.

Models of flow, sediment transport, and bed geometry have been developed for equilibrium situations in bed load and transitional channels with reasonably regular topography (reviewed in Bridge 2003, pp. 193–202), but models of channel evolution by erosion and sedimentation remain on the frontier of 2-D modeling.

Labile Channels

Labile channels are ones that experience sufficiently high values of $\tau*$ that full mobility is frequently experienced over at least some portion of the channel bed. $\tau* > 1$ implies fine sediment that moves readily in suspension. Gravels do not occur in this regime except in exceptionally deep and relatively steep flows ($S > 10^{-3}$; $d > 10$ m), which usually are found only in breakout floods and, perhaps, where major rivers flow in gorges. At $S = 10^{-3}$, $d = 3$ m, the limit sediment size remaining on the bed is 0.18 mm. Sediments finer than approximately 0.18 mm move directly into suspension when entrained (the value may vary somewhat depending on mineral density and turbulence intensity) and are not normally found in significant quantities in the unflocculated state on river beds with velocities of order 1 m s^{-1}, but they do occur in sheltered and quiet waters. Hence, many labile channels are low-gradient channels with hydraulically smooth flows. Such channels typically have $S \leq 10^{-4}$. The bed consists of medium fine to fine sand and banks consist of fine sand to silt, lending them cohesion that is usually reinforced by vegetation to the depth of root penetration. Hence, banks are strong, often sufficiently strong to constrain lateral deformation.

Channels in fine-grained alluvium often exhibit highly sinuous serpentine meanders that grow by loop extension to the point of looping cutoff. Why they exhibit this extreme behavior has not been thoroughly analyzed but is likely related to the ability of the river to suspend nearly all the sediment load, even on very low gradients, so that vertical construction of the inner, prograding meander bank is rapid and early chute cutoff is prevented. Loop extension is a slow process simply because bank strength militates against rapid migration of the channel.

Because sedimentation occurs primarily from suspension, vertical accretion dominates bar and floodplain formation (Taylor & Woodyer 1978, floodplain class C in Nanson & Croke 1992) and the channels are narrow and deep. Channels in fine-grained sediment are often associated with patterns of channel division around long-lived channel islands, a process referred to as anastomosis (type I anabranching channels in Nanson & Knighton 1996). Such channels are common in deltaic settings and in inland basins. The development and persistence of islands undoubtedly is promoted by the dominance of vertical accretion, but appears fundamentally to be caused by avulsion and by the ability of perennial vegetation to become established within the channel zone. Once established, vertical accretion accelerates within and in the lee of the vegetated zone.

THE INVERSE PROBLEM

Basic Relations

Sediment continuity provides the basis for estimating sediment transfer from changes in river channel morphology:

$$\partial q_b/\partial x + \partial q_b/\partial y + (1-p)\partial z/\partial t + \partial C_b/\partial t = 0, \tag{3}$$

in which q_b is bed material transport per unit width of channel, p is the porosity of the sediment deposits, C_b is the concentration per unit bed area of sediment in motion, x and y indicate the downstream and lateral directions, and z is bed elevation. Along a streamline, which may be of central interest in some numerical schemes, the equation can be specified as

$$\partial q_{bx}/\partial x + \partial/\partial x[k_y(\partial C_b/\partial x)] + (1-p)\partial z/\partial t + \partial C_b/\partial t = 0, \tag{3a}$$

in which the first term gives the change in transport with distance along the streamline, the second gives the effect of lateral changes in transport—that is, the effect of sediment lateral diffusion, in which k_y is the lateral sediment diffusion coefficient—the third term gives the deposition, or erosion, and the last term gives the change in sediment concentration in time. Solutions based on field data usually are restricted to timescales much longer than those for synoptic changes in transport, so processes are treated as if effectively averaged and the last term is dropped.

The problem has usually been reduced to one dimension by averaging cross-sectional changes (e.g., Griffiths 1979), whence

$$\partial Q_b/\partial x + (1-p)\partial A_b/\partial t = 0, \tag{3b}$$

in which $A_b = wdz$ is the net deposition or scour in the cross-section of width w.

As a further step, flow equations and a sediment transport function may be introduced to estimate Q_b and a finite numerical scheme employed to find model solutions for sedimentation, as in the well-known HEC-6 program (USACE 1990; see also 1-D programs by van Niekerk et al. 1992, Hoey & Ferguson 1994, Cui et al. 1996, and a review for 2-D and 3-D modeling by Mosselman 2005). Unsteady flow is approximated by steps of steady flow. Today, a number of 2-D codes have been developed but there is as yet no thorough summary of experience available in the open literature.

Inverse estimate: calculation of processes that created a particular system state from observations of that state

Sediment virtual velocity (v_b): the mean rate of travel of a sediment grain between successive surveys, separated by time, T. Hence, $v_b = L/T$. The definition therefore includes the time during which the grain is at rest

The critical importance of inverse estimates of sediment transport and sedimentation from changes in river channel morphology is that they open the way to study variations in transport along the channel, which lie at the heart of understanding how river channel morphology develops. This information is not practically accessible from conventional measurements of sediment transport, which are taken at a single cross-section.

Bedform Migration

Early attempts to estimate sediment transport from morphological changes focused on bedform migration. Exner (1925, in Raudkivi 1990) began from the 1-D form of Equation 3b:

$$\partial q_b/\partial x + (1-p)\partial z/\partial t = 0, \tag{3c}$$

(the so-called Exner equation) and developed the equation for a traveling wave. Simons et al. (1965) integrated Equation 3c into the form

$$q_b = (1-p)v_b\langle h\rangle dw + C, \tag{4}$$

in which v_b is bedform migration rate, $\langle h\rangle$ is bedform average height ($h_b/2$ for triangular dunes of height h_b), and dw denotes unit width. C is a constant of integration to account for the passage of bed material not associated with the migrating bedform (see also Willis & Kennedy 1977). In practice, $\langle h\rangle$ is usually represented as βh_b, in which β is a coefficient for average cross-sectional area of the bedform. For dunes, $0.55 \leq \beta \leq 0.6$, indicating that dunes are somewhat more convex than triangular. Hubbell (1964) made a thorough analysis of the method in historical context, and critical reviews of the method and some variants are given by van den Berg (1987) and ten Brinke et al. (1999). An unusual early demonstration of the method was presented by Wittman (1927), who estimated the sediment transport associated with displacement of gravel bars in a rectified reach of the River Rhine to be approximately 250,000 $m^3\ a^{-1}$, based on nine years of surveys of bar progression.

Critical questions associated with this method are, What is the magnitude of C?, and How can it be determined? Ashmore & Church (1998), reanalyzing the experimental data of Simons et al. (1965), concluded that the waveform solution represented the total sand transport down to approximately 0.28 mm, and was not strongly biased even at 0.19 mm—that is, bed material sand appeared to be substantially entirely associated with bedform migration, whence $C \approx 0$. Of course, this outcome may vary with hydraulic conditions. Dinehart (1992) demonstrated a similar result for rarely observed gravel waves, and an opportunity to compare estimates from sand wave migration with direct bed load–transport measurements (Villard & Church 2003) yielded good correspondence. There remains a need for more critical study of these questions.

Channel Deformation

Rewriting Equation 3b in finite difference form yields

$$\Delta Q_b/\Delta x + (1-p)\Delta A_b/\Delta t = 0, \tag{3d}$$

i.e.,

$$(1-p)\Delta V + (Q_{bo} - Q_{bi})\Delta t = 0 \tag{5}$$

$$\Delta V = V_i - V_o \tag{5a}$$

for an arbitrary time period, wherein V_i and V_o are the volumetric sediment input and output, respectively, for a defined area of the channel bed, and $\Delta V = \Delta A_b \Delta x$ is net change in the volume within a reach of length Δx measured along the channel (**Figure 8**). (By convention, Q_b is represented as mineral volume, whereas deposited or eroded volumes are measured as bulk quantities.) These equations represent the sediment balance for a reach with no significant tributaries. To obtain the sediment transport, one additional condition must be known. Most obviously, this would be the sediment influx at the upstream end of the reach (or efflux at the bottom end). Alternatively, if sediment path length (practically, the mean distance traveled by mobile sediment particles during the measurement interval; see next section) is known, an estimate of sediment transport is accessible.

Information of volume changes along a channel has customarily been obtained from cross-section surveys or from topographic mapping of the channel bed (see, for example, Carson & Griffiths 1989, Lane et al. 1995). The practicality of the technique, then, strongly depends on survey technology. But it also to some degree depends on channel deformation style.

I.V. Popov (1962a,b) is generally credited with having introduced the concept that sediment transport may be estimated from measurement of the sediment balance within a river reach, and he certainly appreciated the need to specify the equations for a particular channel deformation style. Neill (1971, 1987) made the first, and most obvious, application of the method to a regularly meandered river. He measured ΔV as the erosion volume around a meander bend and he assumed that the sediment travel distance was $\lambda/2$, wherein λ is meander wavelength—that is, the sediment proceeds from point of erosion to point of deposition on the next downstream bar (**Figure 8**). A somewhat more general approach uses cross-section surveys to estimate erosion and deposition volumes (Griffiths 1979, Martin & Church 1995). Most generally, digital

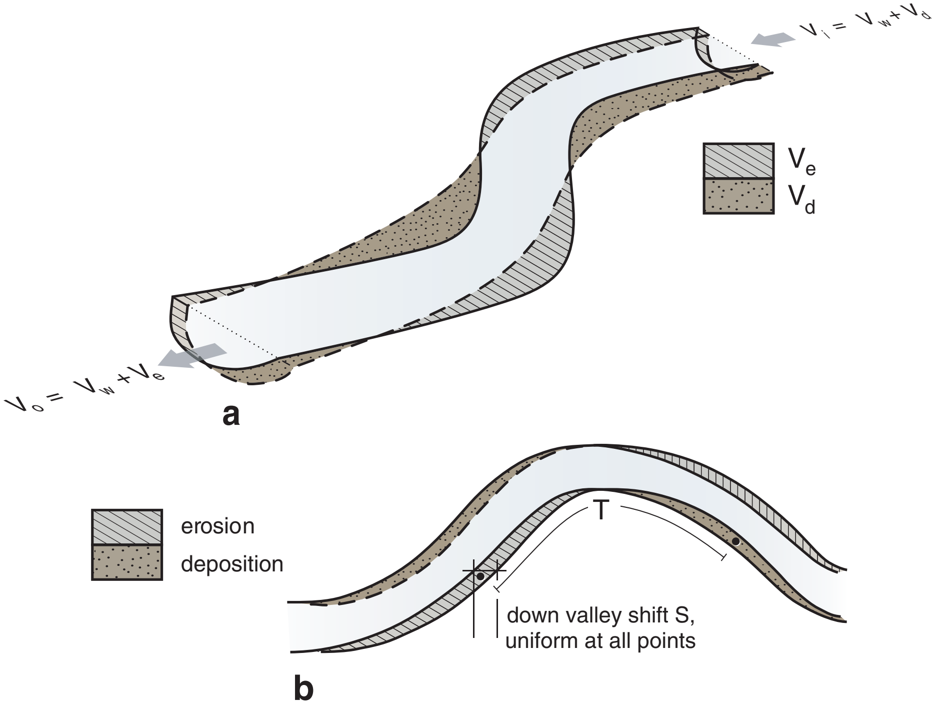

Figure 8

Definition of the sediment balance of a river channel reach, and specification for a regularly meandered channel. V_w is the volume of wash material.

models of elevation difference between successive topographic surveys of the river bed are applied to determine volume change (Lane et al. 1995, McLean & Church 1999).

There are important technical constraints associated with estimates of sediment transport and channel change from changing morphology (Ashmore & Church 1998). Spacing of cross-section surveys (Lane et al. 2003) and point density in distributed surveys (Lane et al. 1994) both influence the realizable precision of estimated morphological change. Compensating scour and fill between surveys introduces negative bias into estimates of change so that results become merely lower-bound estimates (Lindsay & Ashmore 2002). Survey frequency therefore must be related to the timescale of the principal sedimentation events in a river. Similarly, determination of an end-point condition introduces the possibility for significant errors because the measurement of bed-material transport remains a difficult exercise.

In recent years, advances in surveying techniques (Chandler et al. 2002, Brasington et al. 2003, Charlton et al. 2003, Lane et al. 2003) have led to increased exploration of methods based on morphological change to estimate bed material transport and to improve understanding of channel deformation and development. There has as yet been little effort to consider river deformation style systematically to select the most appropriate techniques for application.

Particle Path Length

This topic is worth attention because knowledge of it liberates the problem of transport estimates from reliance on the need for a direct measurement of transport (e.g., Neill 1987). It is also the basis of some independent methods of estimating sediment transport. Path length is the distance traversed by a sediment grain from initial mobilization to final deposition. It may be integrated over several sediment transporting events and may be made up of several individual steps. The practical definition given in the last section equates the time from mobilization to deposition with the intersurvey interval. If particle path length is known and, in addition, the depth of the active layer is known, then sediment transport can be estimated as

Particle path length (L): the distance traveled by a sediment grain between successive surveys; ideally the distance traveled during a single sediment transporting event. Path length may consist of the sum of several individual step displacements

$$Q_b = v_b d_s w_s (1 - p)\rho_s, \tag{6}$$

in which $v_b = L/T$ is the virtual velocity of the particles (i.e., the mean rate of travel, rest periods included), L is clast path length, T is elapsed time between observations of particle position, d_s is scour depth (in certain circumstances, this may be equivalent to z_b), and w_s is the active width of channel bed (that is, the width over which transport occurs). Scour depth might be determined by using scour chains (see Laronne et al. 1994) or tracer clasts (see Hassan & Ergenzinger 2003 for a review; see Sear et al. 2003 for an assessment of scour depth distributions).

It is likely that typical particle path lengths exist in rivers and that they can be inferred from river morphology. Particles accumulate in certain areas, most obviously bars, which is what lends topographic variety to the river bed. It appears likely, then, that average travel distance for a significant transporting event is equal to bar-to-bar spacing, as was in effect assumed by Neill and also by McLean & Church (1999). Direct information derives from studies using tracer stones. Most studies have suggested that path length distributions are positively skewed, with most clasts moving not far at all (see Pyrce & Ashmore 2003a for a comprehensive review), but it also appears that most of the observations have been taken at relatively low flows or in threshold channels that do not exhibit strong sedimentary organization beyond grain scale. In a sequence of laboratory experiments, Pyrce & Ashmore (2003b) have demonstrated that, at dominant or channel-forming flows (in their experiments, runs with $\tau * \geq 0.084$ and $0.1 \leq D_{50}/d \leq 0.15$), clasts indeed tend to travel to and congregate in bars, so that, in the long term, bar spacing is apt to be the dominant path length. More observations are needed, but this conclusion appears to represent an important regularity in the relation between sediment transport and river morphology.

SEDIMENT TRANSPORT PROCESSES AND CHANNEL MORPHOLOGY

Applying Inverse Methods

We have classified three broad types of river channels according to the frequency and mechanics of movement and sedimentation process associated with river bed materials (**Table 1**). Inverse approaches entail making estimates about details of the sediment transport process from the evidence of the deposits and morphology. Such details include volume transported, distance of movement, and the distribution of erosion and sedimentation along the channel. A range of methods is available, and each has certain advantages and disadvantages in relation to river channel type.

Threshold channels are characterized by episodic movement of individual gravel or coarser clasts at relatively low excess shear forces. Scour is not deep and sediment deposits are not stacked (except over the long term if there is persistent aggradation). Consequently, individual clast displacements and net lateral deformation dominate channel changes. In the smallest threshold channels, even limited scour and fill may be important, whereas bank strength is often sufficient to prevent lateral deformation of the channel. In these circumstances, there remains little alternative to direct field survey or to the use of indicators such as scour chains or tracers to discover the changes.

In larger channels, changes can be identified on aerial images from which topographic maps may be prepared. The frequency with which maps must be prepared will depend on the activity of the river and the persistence of sedimentation trends. Braided channels have received major attention (e.g., Lane et al. 2003) because the multiplicity of individual channels, and their instability, makes it extremely difficult to obtain estimates of sediment transport by direct measurements. Methods have been developed to complete mapping through shallow, clear water (Westaway et al. 2000), so that complete maps may be constructed from low-water images. Depths of approximately 0.8 m may be penetrated with good fidelity.

In channels with relatively simple channel geometry, simpler methods may be adopted under certain assumptions. If a characteristic erosion depth can be estimated, then sediment budget estimates can be based on observed planimetric displacement of the channel multiplied by the characteristic erosion depth. In combination with a characteristic sediment path length, this provides sufficient information to estimate sediment transport. McLean & Church (1999) evaluated this method with good results. It is significant because a great deal of historical information might be recovered about period-averaged sediment budgets using map and air photo archives, but such information could be recovered only from rivers with a lateral style of instability. Leys & Werritty (1999) have discussed semiautomated methods for collating information about river channel changes from historical sources.

In transitional channels, both lateral and vertical deformation (scour and fill not directly associated with lateral displacement of the channel) may be important. Leopold (1992) observed that a modest fraction of coarse sediment in the channel bed material dominates the morphology of the channel, and this is implicit in the fractions of bed load versus suspended load in the classification of Dade & Friend (1998) and in **Table 1**. Hence, application of methods based on aerial survey is apt still to be useful. Recently Gaeuman et al. (2003) have constructed a gravel budget for a transitional channel using a combination of estimated travel distances and cross-section surveys. McLean & Church (1999) extracted a gravel budget for a large transitional river in which the throughput sand load is two orders of magnitude larger than the gravel load. The gravel nevertheless dominates the channel morphology, permitting analysis of along-channel variations in gravel transport and deposition. Because of the possibility for deep scour, however, the survey of such channels—if they are of any size—is apt to require boat-borne acoustic sounding, which is expensive to achieve with the strict navigation control necessary for assimilation into the total data set.

The estimation of total sediment transport in such reaches is difficult because the magnitude of throughput wash load—a small fraction of which is deposited on bar tops and on floodplain surfaces—dwarfs the magnitude of the bed material load. Hence, small errors in estimating that portion of the load involved in upper bank sediment exchanges may lead to large errors in the sediment budget and the transport estimates. In some channels dune or sandbar propagation is associated with short-period scour/fill that biases transport estimates based on extended intervals between surveys.

In labile channels bed material sediment budgets cannot, in general, be studied by aerial survey methods. Compensating scour and fill associated with sandy bedforms are a regular feature of sand transport in such channels. We have reviewed methods by which dune propagation might be linked to bed material transport estimates, but in rivers with a high proportion of fine-grained sediment in the channel perimeter, much of the bed material moves in suspension. Nonetheless, methods based on the sediment budget are still accessible provided that records of sediment transport are available at reach limits and at tributary junctions. Direct measurements of suspended sediment are more routine than measurements of bed load, so this may be a viable strategy. A notable attempt to take advantage of sediment transport measurements to construct reach sediment budgets was made by Andrews (1986), who studied the disposition of sand through the Green River system after regulation induced significant aggradation. However, the positions of the gauging stations restricted the computational units to rather long reaches. It is possible that long-term partial sediment budgets may be constructed on the basis of meander migration over periods sufficient to average the effect of dune movements. In such cases, the morphologically significant sediment transfers can be recorded even though the total bed material transport may not.

Current Problems

The foregoing review has been constructed as an attempt to synthesize information about relations between sediment transport and river channel morphology. It reveals some regularity in sediment transport processes, which may help improve our understanding of river processes and practical sediment transport computations. In particular, inverse methods for estimating the sediment transport escape the constraining concept, inherent in classical sediment transport formulations, that rivers must move sediment according to some supposed hydraulic capacity. More significantly, they open the possibility to study the effect of varying transport along the channel, which is the key to understanding morphological variations through time and along river channels.

However, the review also reveals a range of topics that require substantial further consideration. Attention to them is apt to yield rapid progress in understanding.

- It is apparent that the subtleties of meaning associated with definitions and methods of sediment transport measurement require closer attention than they have been given. The customary definitions based on the mechanics of transport—while useful—are not sufficient to understand river channel deformation. Indeed, they may have contributed to limited understanding.

- More needs to be known about the mode of sediment transfer in different channel types; for example, the widespread notion that bed load moves dominantly as a continuous traction carpet in the deep channel appears more to be an artifact of simplified models for computation than a reality of rivers—if it were true, it would completely undermine the basis for inverse estimates of sediment transport.
- The distribution of path lengths in transport of sediments of various sizes urgently requires further investigation.
- More needs to be learned about sediment deposition in relation to river channel deformation and river channel morphology. There is a reasonable understanding of the mechanisms by which sedimentary bedding is created, but not of the ways in which deposits are cumulated into the major morphological features—particularly bars—that define the channel.
- Information is needed on the timescales for significant sedimentation events—that is, for significant river channel change—in different river channel types (see Hoey 1992, for one approach). Such information is needed to constrain the frequency of surveys for the purposes of establishing the sediment budget and understanding river channel deformation.
- More attention must be paid in sediment transport and sediment budget studies to the precision of measurements because precision constrains the interpretations that may be made of river sediment budgets.

CONCLUSIONS

The morphology of an alluvial river channel is the consequence of sediment transport and deposition by the river. It depends in large measure—but not exclusively—on the transport of bed material, that portion of the transported sediments that constitutes the bed and lower banks of the channel. This material may move in traction or, in the sand range, in suspension, so that there is sometimes an ambiguous association between customary sediment transport measurements and the role of the sediments in the channel. Perhaps for this reason, the connections between transport and river morphology have been less closely analyzed than they should.

In this review, a classification, provided by S.A. Schumm, of the relations between sediment transport and channel morphology has been elaborated by examining sediment transport regimes on the basis of the Shields number of the flow—a nondimensional measure of flow forces imposed on the bed that is scaled to sediment caliber. Threshold, transitional, and labile transport regime types are similar to Schumm's bed load, mixed, and suspended load channels, respectively, but additional variations are recognized on the bases of transport rate and the quantity and importance of fine sediment transport, which influences upper bank morphology and the overall morphological expression of the river. These variations, and the entire class of transitional channels, require far more study. Indeed, the threshold channels are the only group for which relatively detailed analyses of relations between transport and consequent morphology are available.

An inverse approach to the problem is introduced, whereby changes in river channel morphology over time are used to back-calculate the bed material transport. For successful application, the method must be adapted to the deformation style of each major transport regime type, lending added significance to appropriate classification. The approach is important because it can reveal the variations in transport along the channel that are responsible for the variations in alluvial morphology and channel deformation. These variations are what make rivers both an interesting and an exceedingly challenging problem to understand.

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