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Dynamic transport capacity in gravel-bed river systems

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Abstract

Sediment transport capacity mediates the transfer and storage of bed material between alluvial reservoirs in a drainage system. At intermediate time scales (10^0 [illegible] yr) corresponding to the evolution of sediment pulses, conditions governing bed-material transport capacity under the hydrologic regime respond to variations in storage and sediment flux as pulses extend, overlap, and attenuate through the network. Relations between transport and storage in degrading gravel-bed channels and general relations between bed-load transport rate and tractive force have similar inflection points marking the inception of armoring. This suggests that in a sediment-charged river where armoring is weak and transport rates are high, transport rates respond weakly to changes in sediment supply, and the channel adjusts primarily by erosion or deposition. At some stage in degradation, armoring commences and responds to variations in sediment input rates, causing sediment transfer rates to decline rapidly as supply and storage decline. However, it appears unlikely that the simple, linear form of transport-storage relations for degrading channels can be extrapolated to full aggradation-degradation episodes: (1) Simulations of transport rates in an initially uniform reach of a natural channel that was affected by an evolving sediment wave show longitudinal variations in relations between bed elevation and sediment flux. (2) In a flume experiment where stepped increases and decreases in feed rate forced episodes of aggradation and degradation, sediment output rates increased after feed rate was reduced and the channel began to degrade. The resulting hysteresis in relations between transport rate and storage was driven by changes in bed texture and channel morphology that increased bed mobility during early stages of degradation before armoring intensified and reduced transport.

Introduction

Understanding the movement of bed material is key to understanding sediment routing and the evolution of alluvial landforms in gravel-bed river systems —even those whose sediment load is dominated by suspended sediment and washload. Along with bank accretion by suspended sediment, erosion and deposition of bed material molds the channel that transfers all sizes of sediment through the system and from channel to floodplain.

Many models of sediment routing and landscape evolution rely on local values of transport capacity to route sediment from one channel segment to the next (e.g., Benda and Dunne, 1997). Given the rate of sediment inputs from upstream and within the segment, transport capacity regulates the net change in storage and transfer to the next downstream segment. The most general definition of transport capacity is the original given by Gilbert (1914): 'the maximum load of a given kind of debris which a given stream can transport'. Recognition that transport capacity is not a fixed quantity but varies with sediment supply and storage (Lisle and Church 2002) leads to a consideration of the time scale over which sediment transport is regarded.

Transport capacity and time scale

Transport capacity at a given point along a drainage network appears constant or dynamic depending on whether the time scale of observation is equal to that of the passage of a sediment pulse or an episode of aggradation and degradation. Transport capacity can be regarded as constant for a time scale corresponding to basin evolution ($\geq 10^3$ yr), since in the long term, the channel must convey sediment at the rate of denudation, taking into account accretion onto internal basins. At this scale, changes in transport and storage associated with the passage of a sediment pulse can be disregarded. Transport capacity can also be regarded as constant at a much shorter time scale corresponding to a hydrologic event ($\leq 10^0$ yr), which is less than that of the passage of a sediment pulse. At this scale, sediment is transported according to the range of flows and the physical conditions of the channel that govern sediment transport. The operative relation between flow and sediment transport represents a snapshot during the evolution of sediment transport conditions as rates of input and storage change. Transport capacity at the time scale of a hydrologic event is not necessarily equal to that representing basin evolution because of variable conditions governing channel mobility, even under a constant hydrologic regime.

Transport capacity is dynamic at an intermediate time scale (10^0–10^2 yr) corresponding to the evolution of a sediment pulse and associated changes in sediment storage in parts of the basin (Lisle and Church 2002). At this scale, conditions governing sediment transport under the hydrologic regime respond to variations in storage and sediment flux as pulses extend, overlap, and attenuate through the network. Sediment pulses are easiest to analyze individually (Nicholas et al., 1995; Lisle et al., 2001), but multiple pulses from tributaries are likely to converge and affect sedimentation zones in a drainage basin as a result of climate change (Macklin et al., 1992; Church et al., 1999), climatic events (Trustrum et al., 1999), major earthquakes (Keefer, 1984), and vegetation changes (Brooks et al., 2003; Trustrum et al., 1999). The pulse scale is most relevant to land management because it spans the progress of sediment-related impacts such as inundation, flooding, and effects on aquatic and riparian ecosystems. If so, understanding the dynamics of transport capacity becomes a critical problem for scientists researching the routing of bed material through drainage systems.

The purpose of the paper by Lisle and Church (2002) was to promote investigation of the relations between sediment transport, supply, and storage. Data on degrading, gravel bed channels from previous field studies and experiments provided an entry to the problem. These showed strong, approximately linear relations between transport rate and volume of sediment stored. In this paper, we examine such variations in transport rate during full cycles of aggradation and degradation, using information from a simulation of sediment-wave evolution and a laboratory experiment. Results indicate more complicated relations than those for degradation alone.

Relations between transport capacity and sediment storage

Armoring, transport intensity, and bed elevation

Lisle and Church (2002) present examples of sediment transport-storage relations for degrading gravel bed channels from flume experiments and two field examples (cited in the article). These examples were chosen to neglect variations in channel gradient and to focus on adjustments in bed armoring. At the beginning of the experiments, armoring was absent or weak because of screeding or high sediment-feed rates. Then, feed rate was cut, discharge was held constant, and sediment output was measured until the bed stabilized. Some experiments exhibited a marked transition in bed conditions as the bed degraded. Initially, armoring was weak, and sediment output rates were high and varied unsystematically about a mean (Phase

I; Figure 1, modified from Figure 2 of Lisle and Church [2002]). Later, armoring intensified and transport rates declined rapidly until the bed stabilized (Phase II). Output rates followed a smooth curve suggesting an exponential decay in sediment storage, or a linear relation between output rate and volume of sediment stored.

The Phase I-II transition in the experiments occured in a range ($0.03<q_s*<0.16$) corresponding to some of the highest transport rates measured in natural channels, including unarmored desert channels (Reid and Laronne, 1995) and channels carrying high loads of sediment from recent volcanic eruptions (Pitlick, 1992**,** Hayes et al., 2002). Dimensionless transport intensity is expressed as $q_s* = \frac{q_s}{\left(gRD^3\right)^{1/2}}$, where q_s is the volumetric transport rate per unit width, g is gravitational acceleration, R is submerged specific gravity of sediment, and D is particle diameter of transported material; dimensionless tractive force (Shields stress) is expressed as $\tau_{50}^* = \tau / R\, D_{50}$, where τ is mean boundary shear stress and D_{50} is the median particle size of the bed surface. An inflection appears in these relations at a value of $q_s*\sim 0.1$ and a value of $\tau_{50}^* \sim 0.1$; beyond this, transport rate increases more gradually with increasing Shields stress **(**Figure 2). In this range of τ_{50}^*, armoring largely disappears (Parker and Klingeman, 1982; Wilcock and Southard, 1989). The transport formula of Wilcock and Crowe (2003) are nonlinear, but show an increase in curvature at approximately the same value. (The agreement of the formula with the data is only fair at high Shields stress. Much of the discrepancy may arise from the tendency for greater abundance of sand on the bed surface at high Shields stress, indicating a shift in appropriate transport relations from sand-poor to sand-rich.) Taken together, these observations tend to confirm experimental evidence that an inflection in transport-tractive force relations occurs at a point where armoring disappears. Below this transition, a mobile armor inhibits bed mobility and makes transport highly sensitive to variations in tractive force. Above this transition, armoring no longer influences bed mobility, and transport increases more gradually with applied stress. Weak armoring is commonly associated with aggraded channels (Lisle and Madej, 1992; Gomez et al., 2001; Hayes et al., 2002) and indicates achieving the upper limit of transport capacity (Dietrich et al., 1989).

The general form of the transport-tractive force relation (Figure 2) resembles the general form of transport-storage relations of degrading gravel-bed channels proposed by Lisle and Church (2002) (Figure 3). The dimensionless transport rates for each channel describe a locus of points along a general transport relation; those from more mobile, sediment-charged channels plotting high along the curve. Considering the importance of high flows in channel evolution, we can characterize sediment-charged channels (e.g., Pasig-Potrero) by Phase I conditions, and sediment-limited channels (e.g., Oak Creek) by Phase II conditions.

The two relations—transport-tractive force and transport-storage—are not only similar in form but also linked by process. In both, an inflection in the relation occurs where armoring essentially disappears. This can be interpreted as follows. In a sediment-charged river where armoring is weak and transport rates are high, transport rates respond weakly to changes in sediment supply, and the channel adjusts primarily by erosion or deposition. At some stage in degradation, armoring commences and responds to variations in sediment input rates, causing sediment transfer rates to decline rapidly as supply and storage decline. The supposition is that with increased armoring, Shields stress at a given discharge would decrease, causing the locus of points on the transport-tractive stress curve to shift downward, given no change in hydrologic regime. This conceptual model has not been tested in a natural system, but Lisle et al. (2000) correlate high reach-averaged, bank-full Shields stress with sediment supply, and Sawada et al. (1985) and Laronne et al. (2001) report shifts in bed-load rating curves with changes in the abundance of gravel and sand on stream beds that are dominated by cobbles and boulders.

A change in transport rates corresponding to a change from a well-armored bed to an unarmored bed can be estimated from common values of Shields stress for armored channels and the Wilcock-Crowe transport function. Andrews (1984) reports an average bank-full Shield stress value of 0.046 for twenty-four rivers in Colorado, which have low bed-material supplies. If we assume that an average degree of armoring (ratio of median sizes of surface material and bed load) equals 4 and allow for hydraulic changes due to bed roughness, bank-full bed-load transport ranging from a well-armored to weakly armored condition would vary over approximately one to two orders of magnitude. This suggests a need to include variations in transport capacity in bed-material routing schemes.

The unarmored condition (Phase I) is traditionally identified as 'transport-limited', meaning that sediment-supply rates equal or exceed a fixed value of transport capacity, and flux rates depend solely on discharge. Dietrich et al. (1989) identify transport capacity as being met when armoring disappears with greater sediment supply. Such a high reference value of transport capacity may be appropriate for landscape evolution models used to predict the construction of alluvial landforms during periods of high sedimentation. However, even in Phase I conditions, a single value of transport capacity may not be sufficiently accurate:

1) Changes in channel morphology commonly accompany aggradation and degradation and influence transport capacity (Lisle, 1982). According to the stable channel model of Parker (1978), an increase in bed material supply causes the central mobile zone of a gravel-bed channel to expand into the foot-slope of the stream banks. As a result, the banks erode, the channel widens, and excess boundary shear stress in the center recovers to the value that coexists with stable banks. Although channel widening would limit maximum unit transport rates, total transport rates would increase. Therefore, estimates of a high reference transport capacity corresponding to the absence of armor must account for potential changes in channel width, as well as channel morphology (e.g., braiding).
2) Bed surfaces in gravel-bed channels are not uniform but commonly patchy (Seal and Paola, 1995; Lisle and Madej, 1992). Variations in bed mobility are manifested as variations in relative areas of bed patches having different surface texture (Lisle et al. 2000), so that even aggraded channels exhibit significant armoring when bed textures are spatially averaged (Lisle and Madej, 1992; Gomez et al., 2001). Therefore, weak armoring would not occur uniformly over the channel but would grow in area as sediment supply increased, thereby increasing transport capacity beyond an initial transition to Phase I conditions locally.
3) Transport capacity is not only influenced by sediment supply but also by structural elements (e.g., large wood) that influence the fraction of the total tractive force that is exerted on bed particles (Smith et al., 1993; Buffington and Montgomery, 1999; Brooks et al., 2003). The routing of large sediment fluxes can be strongly influenced by the presence of wood (Swanson and Lienkaemper, 1978; Montgomery et al., 2003**,** May and Gresswell, 2003).
4) Significant changes in gradients of alluvial reservoirs are possible during the growth, decline, and interaction of sediment waves. Gradient is a dominating factor in transport capacity (Gilbert, 1914).
5) Inputs of sediment can change the particle size of bed material. The particle size of the bulk of the input may be different than that of the ambient bed material, and sorting of the input after it enters the channel can deliver a finer component to downstream reaches.

The armored condition (Phase II) is traditionally identified as 'supply-limited', meaning that sediment-supply rates do not meet the transport capacity of the unarmored condition. Lisle and Church (2002) contend that changes in supply during Phase II conditions cause changes in both transport and storage. If so, transport capacity is always adjusting to the rate of sediment supply, with changes in storage as an outcome. An important threshold for basin evolution is reached when sediment storage decreases to the stage that bedrock is exposed to channel erosion. At this stage, bedrock exposure and erosion are governed by sediment supply and transport capacity (Sklar and Dietrich, 1998). Evaluating the transport capacity at this stage (presumably much less than that in Phase I conditions) could be important for parameterizing a model of basin evolution.

Sediment transport over a sediment wave

Changes in sediment storage and associated variations in sediment transport rate that are observed in a channel/valley segment can be considered to be the outcome of the extension and dispersion of a bed-material wave. Once emplaced by a large input, individual bed-material waves in characteristic, quasi-uniform, gravel-bed channels disperse with negligible translation (Lisle et al., 2001). Dispersion of a sediment wave involves the stretching of a downstream-fining sequence created by selective transport interacting with an evolving longitudinal profile (Paola et al., 1992; Cui et al., 1999). Results reported in the preceding citations indicate that the downstream variation in median particle size, for example, is constant when downstream distance is normalized by the total length of the wave and particle abrasion is neglected. Extension of the wave forces at-a-section changes in channel gradient, bed-material size, and channel morphology. This concept predicts thatthe sediment transported through a reach and composing the active bed material would become finer with the arrival of the leading edge of the wave and then coarsen as the sediment stage rises and later falls with further attenuation of the wave. Textural changes in sediment supply alone would argue for a rapid rise in bed mobility as the wave extended into a reach, followed by decreases in mobility, even before peak aggradation occurred. However, transport rates would also be mediated by the rate of sediment supply from upstream, which would also affect bed mobility.

There are no detailed records of sediment transport rates along the path of a sediment pulse, but general patterns can be simulated using the sediment-wave model of Cui and Parker (in press) applied to a natural channel where the model predictions are accurate (Sutherland et al., 2002). The wave originated in March, 1995, from a landslide that delivered 60,000 m^3 of mixed-size sediment and buried the channel of the Navarro River in northern California by as much as 9 m. As the wave evolved over the next 5 years, the wave crest remained stationary at the point of entry of the landslide and eroded to within 1 m of the original bed elevation. Measurable deposition occurred as much as 2 km upstream and downstream of the original deposit as the wave dispersed.

Annual topographic surveys, sediment sampling, and daily mean discharges that were measured at a nearby gauging station were used to test the accuracy of predictions of the Cui-Parker model. The model accurately hindcasted general changes in bed elevations (Sutherland et al., 2002), suggesting that local transport rates can be accurately simulated. Figure 4A shows longitudinal variations in transversely averaged bed elevations and local bed-load transport rates from the wave crest downstream. Degradation occurred in the upstream limb of this segment (2.5-3.2 km) during the first four years, while the downstream limb aggraded slightly and then degraded. Local transport rates were computed for recorded discharges of similar magnitude (130-180 m^3s^{-1}) that were selected to represent each water year. At each location, bed elevation and transport rate appear to be positively correlated, as the rankings of the two variables for each year are consistent. However, relations between transport rate and bed elevation (de-trended using the estimated original profile) are not consistent between locations (Figure 4B). Because the study reach was approximately uniform, this indicates that the sedimentology and morphology of the original landslide deposit locally controlled transport-storage relations in the affected reach. This suggests that a different deposit would have produced different sediment-storage relations at the same locations. If so, transport-storage relations could not be accurately predicted by the sedimentology and morphology of the pre-existing channel alone, but would also depend on the sedimentology, morphology, and location of the imposed sediment input.

Transport-storage relations during full episodes of aggradation: a flume experiment

We conducted a flume experiment to investigate relations between sediment transport and storage during sediment pulses that created cycles of aggradation and degradation. We modeled Cuneo Creek, a tributary of Bull Creek flowing into South Fork Eel River, northern California. Cuneo Creek received two major sediment pulses in 1955 and 1964 that caused total aggradation of approximately [illegible]0 m (Short, 1987). The channel has since degraded to within 1-2 m of its pre-flood elevation. Experimental conditions are given in Table 1. Water discharge was held constant and two sediment pulse events (Figure 5A) were induced according to the following procedure:

1) Feed sediment at a low rate (4.0 g/s) until equilibrium is achieved.

2) First pulse: Increase the feed rate to 22.4 g/s and hold it steady as the channel aggrades, then reduce feed rate to an intermediate value (12.0 g/s) before equilibrium is reached.
3) Second pulse: Repeat step 2, using the same feed rates.
4) Final: Maintain intermediate feed rate until equilibrium is reached. Then reduce rate further to 6 g/s and allow equilibrium to be established.

We measured sediment output rate continuously, and measured surface particle size and channel topography periodically during the experiment.

Table 1. Experimental conditions

Parameter	**Value**
Flume Length, width	12 m, 0.76 m
Slope	3%
Discharge	1.1 l/s
Scale Factor	25
Sediment Distribution	Range = 0.25 – 11.2 mm Median = 1 mm Standard Deviation = 1.8 mm

The initial channel was well armored and contained a single-thread, alternate bar sequence. Both pulse events showed a similar sequence of changes that progressed downstream with each change in feed rate. When feed rate was increased, the channel became covered with moving particles and armoring disappeared. While fine particles blanketed most of the bed, small congested zones of large particles formed and broke up. The active bed widened, and the channel later became braided. During aggradation, output rates increased but did not equal the input rate (Figure 5B**)**. Therefore, equilibrium was not achieved before input was reduced and the channel began to degrade, as was probably the case in Cuneo Creek. When feed rate was reduced, output rates increased further. A single-thread channel was incised and was covered by rapidly moving, mostly fine particles. Later, the channel armored as the zone of active transport contracted and output rates decreased to equal the intermediate feed rate. Changes in transport rate and bed texture mimicked Phase I/Phase II transitions observed in previous degradation experiments, i.e., high, variable transport in an unarmored bed was followed by armoring and steadily decreasing transport rates. Armoring progressed further after the final feed-rate reduction.

Transport storage-relations during full sequences of aggradation and degradation were more complex than those observed during previous degradation experiments (Figure 5C). Transport rates were lower during aggradation than during Phase I degradation, creating strong counter-clockwise hysteresis in transport-storage relations. Some hysteresis is the inevitable result of the lag in output to changes in input at the head of the flume. However, we believe that some is due also to changes in transport efficiency. During aggradation, clusters of large particles formed intermittently over most of the bed and prevented the integration of smooth zones (Iseya and Ikeda, 1987) that could efficiently transfer all sediment sizes. During degradation, many large particles came to rest, allowing finer particles to become winnowed and concentrate into extensive smooth zones. Moreover, incision increased flow depth, thereby increasing boundary shear stress acting on a bed that remained unarmored.

The bed stabilized at a higher elevation after the second degradation than it did after the first. The increase in apparent base level was due to the accumulation of large particles during each cycle of degradation.

Conclusions

The recent paper by Lisle and Church (2002) introduces transport-storage relations as a means to improve sediment routing models. While there is convincing evidence that transport capacity responds to changes in sediment supply and storage, there is little information on the form of transport-storage relations other than that they are generally positive. Degrading gravel-bed channels provide the best documented examples of transport-storage relations. They show an approximate exponential decrease in storage with time (disregarding hydrologic variation), which implies a linear relation between transport capacity and storage. Underlying this relation is high sediment transport and a lack of armoring at a high sediment stage, followed by the onset of armoring and declining transport rates.

However, our investigations using limited data suggest that transport-storage relations for degrading alluvial reservoirs cannot be extrapolated accurately to full cycles of aggradation and degradation. Variations in transport rate over a sediment wave were simulated using a model of wave evolution (Cui and Parker, in press) that successfully predicted annual changes in bed elevation in a natural channel during a 5-year period. Different locations along the path of the wave exhibited different relations between bed elevation and transport rate, despite the uniformity of the channel before the wave was introduced. Sediment sorting and morphologic changes imposed by the spread of the wave created longitudinal variations in relations between bed elevation and sediment flux. A flume experiment using imposed sediment pulses showed strong hysteresis in sediment output rate and volume of sediment stored: Output rates during aggradation were lower than those during degradation partly because of different particle-particle interactions and channel morphology. Moreover, hysteretic relations for two pulses described different patterns.

These results, as well as other untested influences on transport capacity, suggest a need for further investigations of the dynamics of bed material routing in a drainage basin. At present, routing models based on the transfer of sediment between alluvial reservoirs according to fixed values of transport capacity would likely be inaccurate at the time scale of the evolution of a sediment pulse. However, accurate models based on dynamic transport capacity according to transport-storage relations appear to be elusive. More research is needed, particularly in natural channels undergoing rises and falls in sediment stage. An initial objective would be to appraise the level of precision needed to model sediment routing in practical applications, given the range of variation of transport capacity under different conditions.

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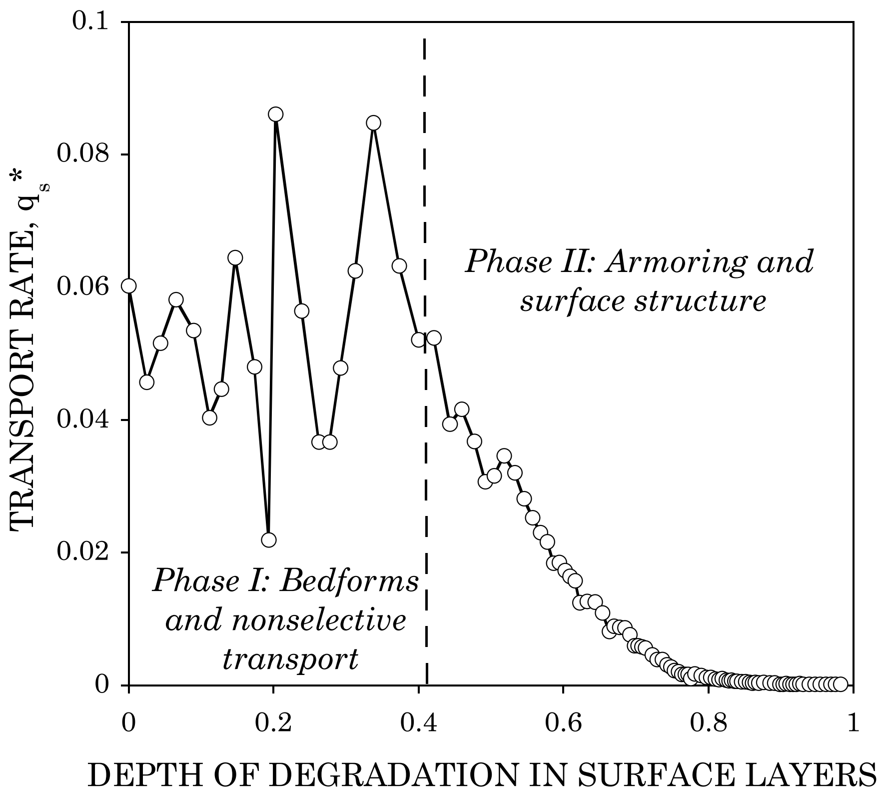


1. Figure 1. Variation of bed-load transport rate and depth of degradation for Run 6.1 of *Little and Mayer* [1972]. Degradation was computed as the average change in bed elevation for the entire flume. Time progresses from left to right. Positions of boundaries between phases of bed-surface evolution were interpreted from the authors' description of the bedform-armoring transition.

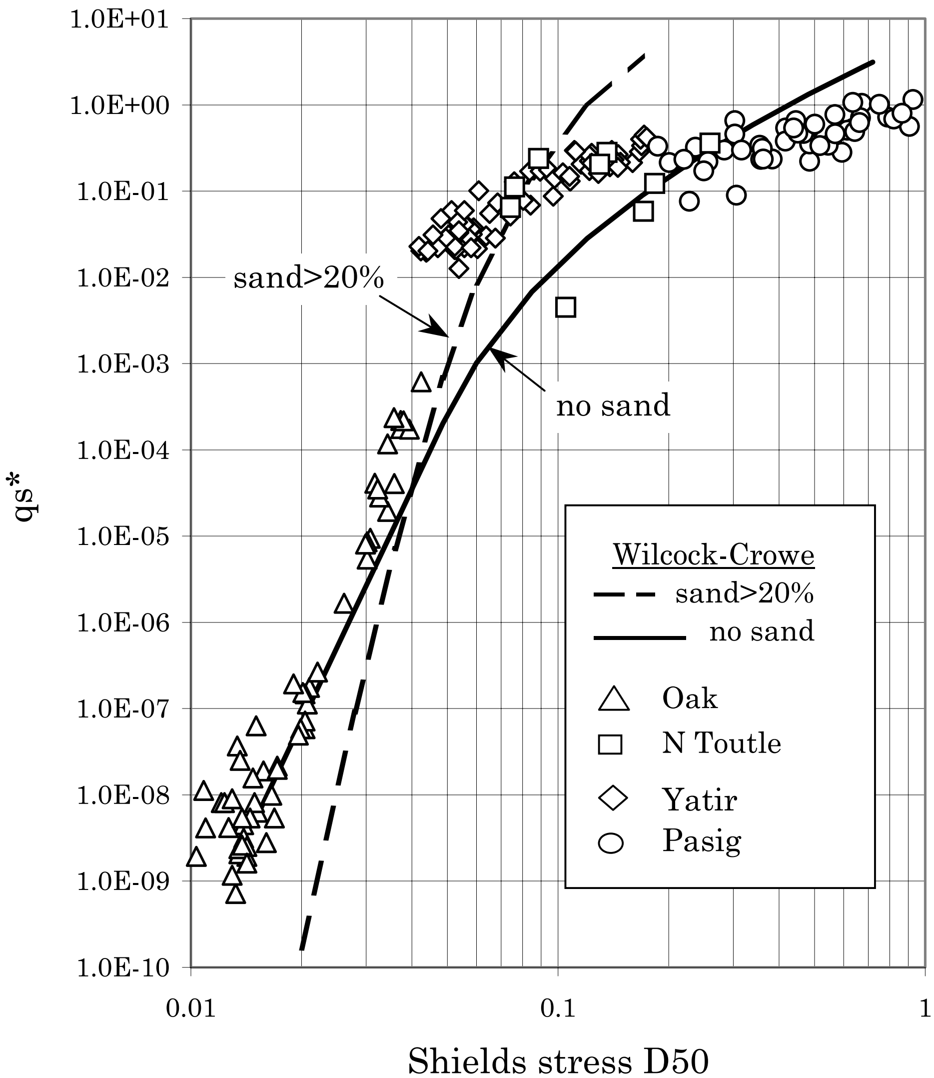
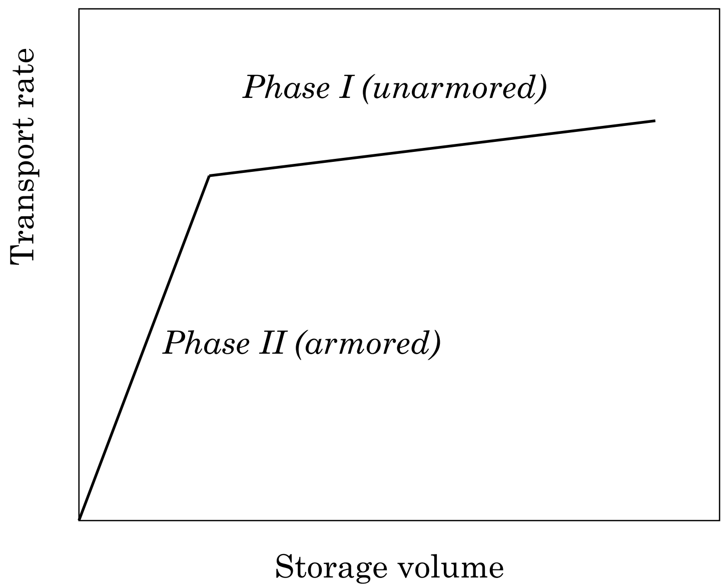


Figure 2. Bed-load transport relations for natural channels and the surface-based function of Wilcock and Crowe (2003).

Figure 3. General transport-storage relation for degrading alluvial reservoirs (adapted from Figure 9 of Lisle and Church [2002]).



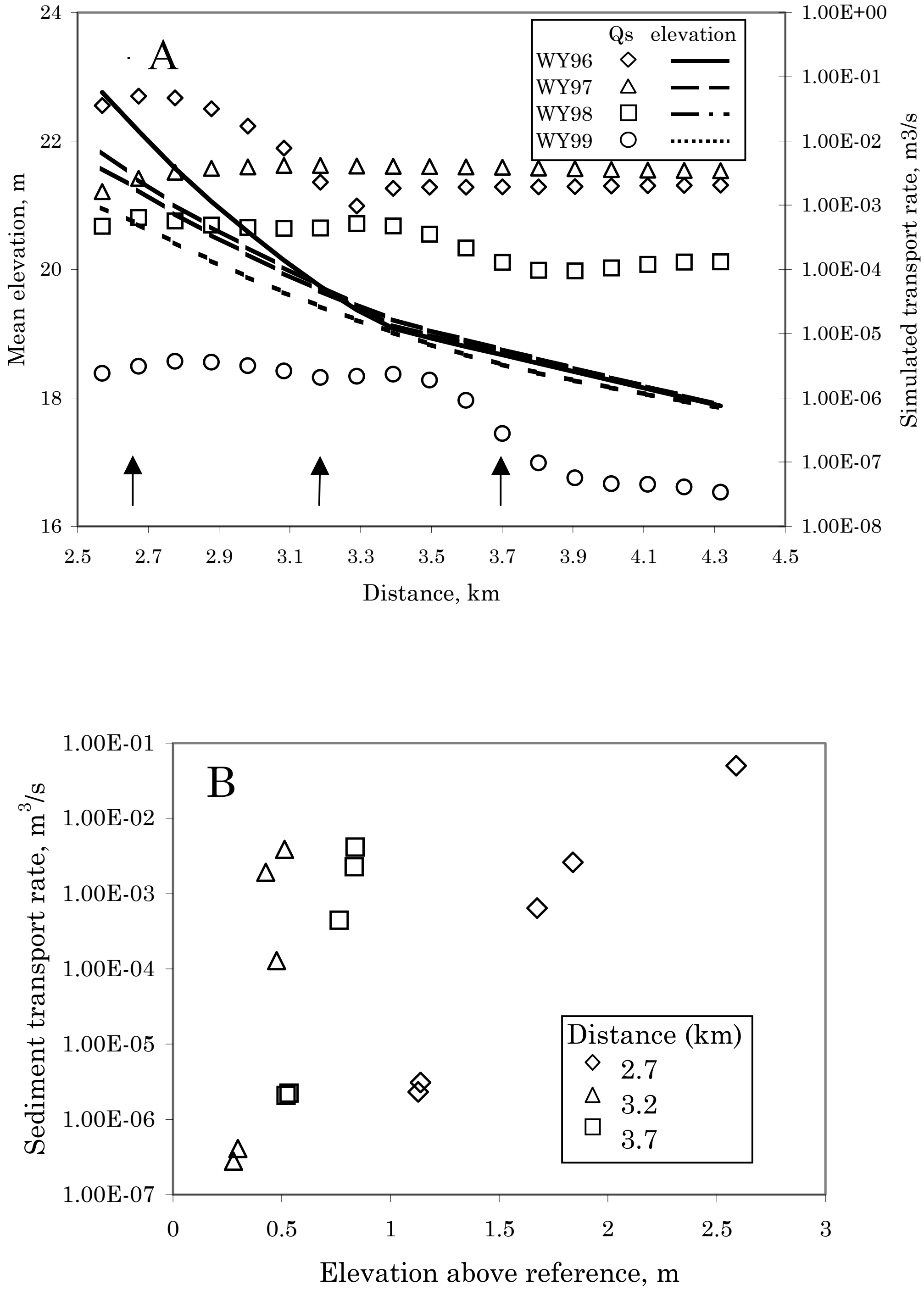


Figure 4. Simulated patterns of bed-load transport rate at discharges of 130-180 m^3s^{-1} downstream of the crest of a sediment wave in the Navarro River, California, for water years 1996-1999: (A) Downstream variation in transversely averaged bed elevation and transport rate; (B) Relations between transport rate and de-trended bed elevation at three locations (shown by arrows in A).

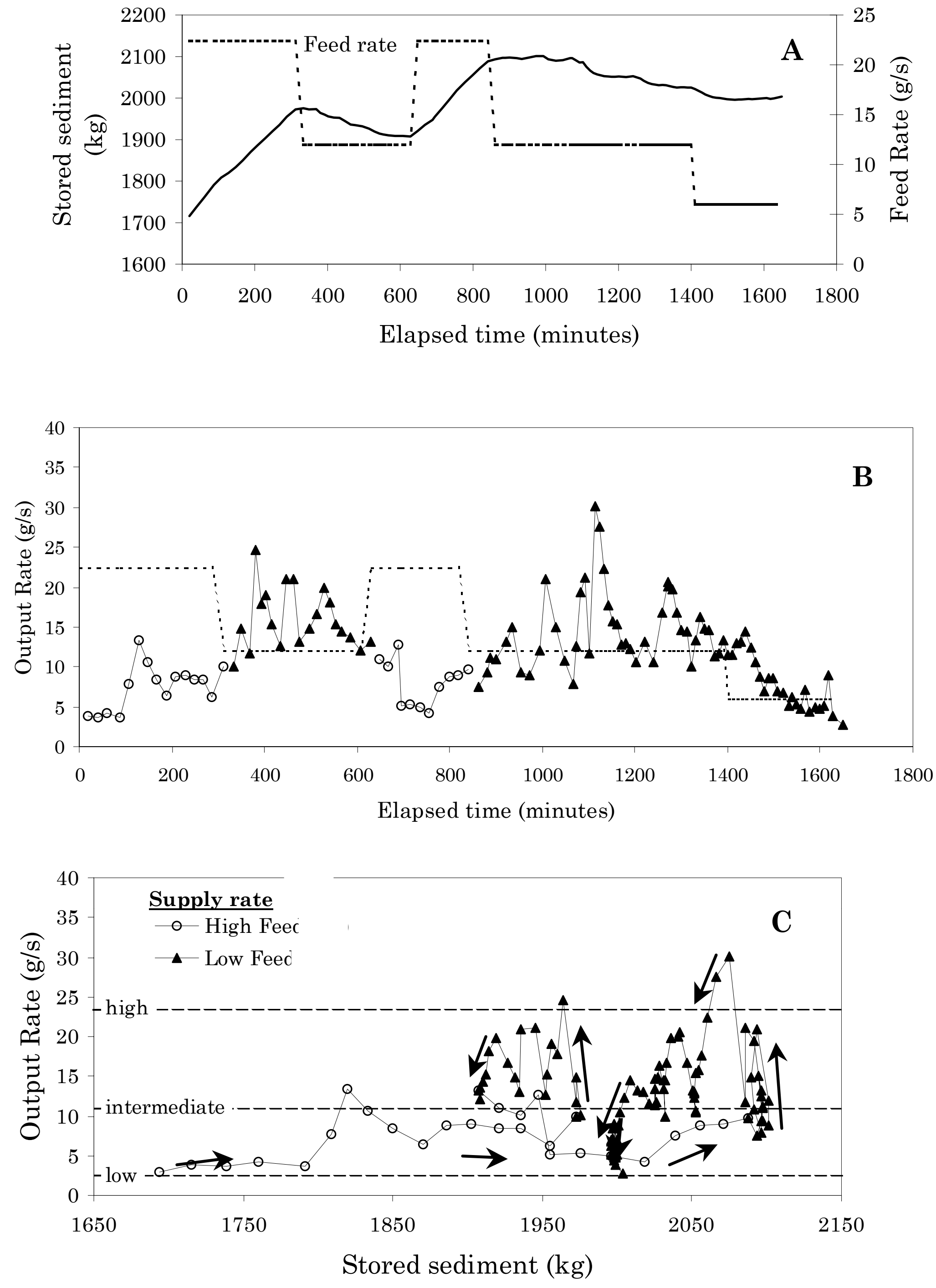


Figure 5. Results of an experiment testing the effects of sediment pulses on bed load transport and storage, showing temporal variations in feed rate and (A) stored sediment and (B) output rate; and (C) output rate *vs* stored sediment.