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Measurement of Bedload Transport in Sand-Bed Rivers: A Look at Two Indirect Sampling Methods

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Abstract

Sand-bed rivers present unique challenges to accurate measurement of the bedload transport rate using the traditional direct sampling methods of direct traps (for example the Helley-Smith bedload sampler). The two major issues are: 1) over sampling of sand transport caused by “mining” of sand due to the flow disturbance induced by the presence of the sampler and 2) clogging of the mesh bag with sand particles reducing the hydraulic efficiency of the sampler. Indirect measurement methods hold promise in that unlike direct methods, no transport-altering flow disturbance near the bed occurs. The bedform velocimetry method utilizes a measure of the bedform geometry and the speed of bedform translation to estimate the bedload transport through mass balance. The bedform velocimetry method is readily applied for the estimation of bedload transport in large sand-bed rivers so long as prominent bedforms are present and the streamflow discharge is steady for long enough to provide sufficient bedform translation between the successive bathymetric data sets. Bedform velocimetry in small sand-bed rivers is often problematic due to rapid variation within the hydrograph.

The bottom-track bias feature of the acoustic Doppler current profiler (ADCP) has been utilized to accurately estimate the virtual velocities of sand-bed rivers. Coupling measurement of the virtual velocity with an accurate determination of the active depth of the streambed sediment movement is another method to measure bedload transport, which will be termed the “virtual velocity” method. Much research remains to develop methods and determine accuracy of the virtual velocity method in small sand-bed rivers.

Introduction

Knowledge of sediment transport is important to such endeavors as river restoration, ecosystem protection, navigation, and infrastructure management. The processes governing sediment transport are complex. As such, accurate observations of sediment transport are crucial to provide data to properly formulate understanding of the sediment transport process.

An alluvial river is a water body that flows through gravels, sands, silts, or clays deposited by flowing water. Natural alluvial rivers are usually wide with an aspect ratio (width to depth) of 10 or greater (Yalin and da Silva, 2001) and the boundary can be molded into various configurations as was demonstrated in the seminal work by Gilbert (1914). With alluvial rivers, the channel geometry is influenced not only by the flow of water but by the sediment transported by the water. When the flow discharge changes, the sediment transport changes and, in turn, the channel geometry usually changes. This channel geometry change, common in sand-bed channels, can then influence changes in the stage; which results in further changes in sediment transport.

Sediment transport is divided into bed-material load and wash load. The bed-material load is defined as that part of the sediment in transport whose sizes are found predominantly in the bed, whereas the wash load is defined as that part of the sediment in transport that is not found predominantly in the bed.

Bed-material load is further divided into the categories of suspended bed-material load and bedload. Sands and gravels are stationary on the bed of an alluvial river until some critical state of flow is reached. Shields (1936) proposed a curve of critical dimensionless shear stresses to delineate stability criteria for particular bed material sizes. There has been much dispute over the values of dimensionless shear stress that will cause particles to move (Garcia, 1999) and whether the critical stress is deterministic, as purported by Shields, or, rather, stochastic, owing to the presence of turbulence as a main mechanism in the entrainment of particles as well as to the hiding factor that accompanies sediment mixtures.

Once the critical state is reached, sand and gravel are transported by skipping and hopping along the bed (known as bedload or saltation load). Furthermore, sand may be entrained up into the water column for a short time period, to be part of the suspended sediment load, and then re-deposited and often returned to bedload. This process is on a sort of continuum cycle whereby the circularity is dependent on the following: flow velocity, turbulence intensity, grain shear stress, entrainment of sediment from bed, concentration of sediment in the water column, sediment size, bedload transport, and the formation and maintenance of bed forms. Understanding this process is key to understanding the behavior of an alluvial river.

The definition of bedload varies around the theme that the particles that are moving stay in close proximity of the bed. Bagnold (1973) concluded that bedload was governed by gravity alone (as opposed to suspended sediment, which has the added process of diffusion) and that the particles stay in successive contact with the bed. Einstein (1950) defined the bedload as being a layer, two sediment diameters thick, with the saltation particles being part of the suspended-sediment load. Vanoni (1975) chose not to separate out the saltation load, as the border between contact and saltation load is not well defined, and defined bedload as “a material moving on or near the bed so that the total load is now made up of the bedload and the suspended load”. A similar definition, and one that is used for this paper, is that of Garcia (1999) who defines bedload transport as that part of the total sediment load in a channel where the particles roll, slide, or saltate over each other, staying in close proximity of the bed.

The flow and sediment transport dynamics of the sand-bed river is further complicated by the various types of bedforms occurring as a result of the interaction between the flow and the erodible bed. The progression of bedforms appears in two regimes of flow: lower flow regime, which occurs during sub-critical flow, and upper flow regime, which occurs as the flow approaches critical to super-critical conditions. Sub- and super-critical flow conditions are defined by the Froude number

$$\left(F = U / \sqrt{gH} \right), \quad [1]$$

where U is the mean velocity, g is the acceleration of gravity, and H is the depth of flow. As the Froude number increases, the lower flow regime bed configuration transitions as follows: ripples, dunes, and washed out dunes (also called transition). The lower flow regime progresses into the upper flow regime as the Froude number increases. The progression and transition of the upper flow regime is: plane bed, antidunes, and chutes and pools.

For most sand-bed rivers of interest, the lower regime will be of primary interest as the Froude numbers are typically small (F<0.3). These small Froude numbers usually preclude the possibility of antidunes, which typically develop only as the Froude number approaches 1.0 (Simons and Richardson, 1966). However, plane beds, which are typically the result of only upper regime processes, can also develop in the lower regime over a large range of Froude numbers (0.3<Fr<0.8).

The transport of bed sediments is dependent on the skin friction between the water and the bed sediments. The formation of bedforms adds an additional component to the total shear stress because of the form drag that occurs in the presence of the bedforms. Einstein (1950) presented the following partition for the components of the shear stress

$$\tau_b = \tau_{sf} + \tau_{fd}, \qquad [2]$$

where τ_b is the total bed shear stress, τ_{sf} is the skin friction shear stress (or grain shear stress) and τ_{fd} is the form drag shear stress. The form resistance stems from the local flow separation and re-circulation in the lee of the dune. This partitioning of the shear stress is important for determining the bedload transport rates, as only the grain shear stress influences the sediment transport. The form shear stress is a result of pressure differences over the bedform (which are normal to the surface) and has no impact on the sediment transport.

Bedload Measurement Methods in Sand-Bed Rivers

Bedload is difficult to measure, especially by direct measurement. Bedload is highly variable in space and time across the river, thus any sampling scheme must take this into account. This variability is exacerbated when bedforms are present. Bedload measurement can be divided into two types: 1) direct measurement with a physical trap and 2) indirect measurement using some sort of remote sensing technique. In this paper, the direct measurement method, as well as two indirect measurement methods are discussed: i) bedform velocimetry and ii) virtual velocity of the bed material.

Bedload Measurement with Physical Trap

Direct measurements of bedload have traditionally been made by placing samplers in contact with the bed, allowing the sediment transported as bedload to accumulate (or be trapped) inside the sampler for a certain amount of time, after which the sampler is raised to the surface and the material is emptied and weighed to determine a weight transported per unit time. Over the years, many types of bedload samplers have been proposed and used. Vanoni (1975) describes some of the more specialized and obscure samplers. Hubbell (1964) characterizes the three types of direct samplers as: 1) box or basket, 2) pan or tray, and 3) pressure difference. The box samplers consist of an open front box or basket, which is lowered to the stream bottom and allows sediment to enter. These samplers have various sampling efficiencies, where sampling efficiency is defined as the ratio of sampled bedload to ambient bedload transport (some would define this as error). Pan or tray samplers consist of a pan or tray that sits out in front of the opening to a box. Again, this type of sampler has varying efficiencies. The most predominant of the direct bedload samplers is the pressure-difference sampler.

The pressure-difference sampler is designed to generate a pressure drop at the exit of the nozzle so that the sampling efficiency is close to 1. The pressure difference is formed by designing a larger ratio of the exit cross-sectional area than the entrance cross-sectional area (area ratio). This difference in cross-sectional area can result in hydraulic efficiencies that are not equal to 1, where hydraulic efficiency is defined as "…the ratio of the mean velocity of flow through the sampler entrance to the mean velocity of the flow through the area occupied by the sampler entrance when the sampler is not present…" (Hubbell and others, 1985). The best-known pressure-difference sampler is the Helley-Smith bedload sampler (figure 1). The Arnhem or Dutch sampler is an earlier pressure-difference sampler on which the Helley-Smith design is based. The original Helley-Smith, weighing 30 kg, had a 7.62-cm by 7.62-cm opening with a ratio of entrance to exit areas (area ratio) of 3.22. Later versions have weighed up to 250 kg for use on the Amazon (Emmett, 1980). The 7.62-cm. Helley-Smith sampler was designed for particle sizes from 2 to 10 mm and with mean velocities up to 3 m/s (Emmett, 1980) with later versions of the Helley-Smith having 15.24-cm by 15.24-cm openings to accommodate larger particle sizes (Edwards and Glysson, 1999). Results of laboratory and field studies on the sampler efficiency of the Helley-Smith vary. Helley and Smith (1971) indicated sampling efficiency of 160 percent, Emmett (1980) concluded from his field study that the sampling efficiency was around 100 percent, while the laboratory study of Hubbell and others (1985) indicated a sampling efficiency of 150 percent for sands and close to 100 percent for gravels. Emmett (1980) states that when the sediment size is near that of the mesh size of the bag (0.2 - 0.25 mm) sample material either escapes or plugs the bag, decreasing the sampler efficiency. Hubbell and others (1985) conducted tests on a modified Helley-Smith sampler with an area ratio of 1.40, which resulted in a sampling efficiency of about 100 percent for all sediment sizes. This new sampler was accepted in 1985 by the Technical Committee on Sediment (now the Federal Interagency Sedimentation Project Technical Committee) as a provisional standard sampler for use by U.S. Federal agencies. The sampler has been designated the USBL-84. The BL-84 has been endorsed by the U. S. Geological Survey (USGS) as the sampler of choice (Edwards and Glysson, 1999), however, Helley-Smith sampler data are still accepted until more testing has been done. Table 1 contains the various versions of the Helley-Smith sampler listing the comparative area ratios and hydraulic efficiencies as taken from Hubbell and others (1985).

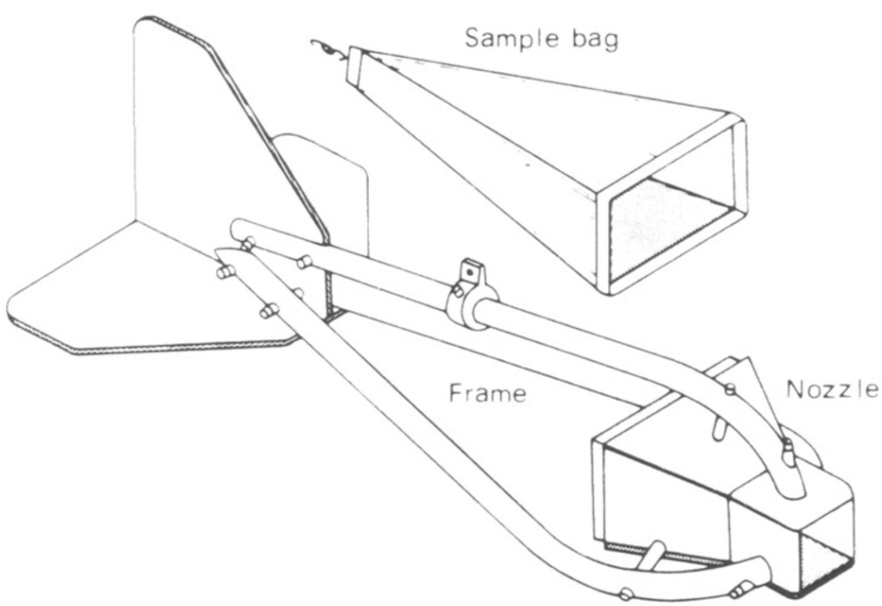

Figure 1. Helley-Smith Bedload Sampler (from Emmett, 1980).

Table 1. Area ratios and hydraulic efficiencies for various Helley-Smith bedload sampler configuations (from Hubbell and others, 1985).

Intake Nozzle Entrance Size[A]	Area Ratio	Hydraulic Effeciency
3 x 3	3.22	1.54
3 x 3	1.10	1.15[B]
3 x 3	1.40	1.35[B]
12 x 6	1.10	1.15[B]
12 X 6	1.40	1.40[B]
6 x 6	3.22	1.54

A. Width x height of the nozzle entrance, in inches

B. Estimated by Hubbell and others (1985)

Bedload Measurement Using Bedform Velocimetry

The two-dimensional Exner equation for bed sediment mass conservation can be written as

$$\left(1-\lambda_p\right)\frac{\partial \eta}{\partial t}+\frac{\partial}{\partial t}\left(hC\right)=-\frac{\partial}{\partial x}\left(q_{bx}+q_{sx}\right)-\frac{\partial}{\partial y}\left(q_{by}+q_{sy}\right), \qquad [3]$$

where λ_p is the porosity of the bed material; η is the local bed level above the datum; C is the depth-averaged suspended-sediment concentration; q_{bx} and q_{by} are the bedload transport in the x and y directions, respectively; and q_{sx} and q_{sy} are the suspended-sediment loads in the x and y directions, respectively. Simons and others (1965), starting from the one-dimensional (streamwise) form of equation 3 and assuming that the suspended sediment was in equilibrium ($dq_s/dx=0$), showed that the volumetric bedload per unit width, q_b, for idealized triangular bedforms could be computed as

$$q_b=\left(1-\lambda_p\right)U_b\frac{\Lambda}{2}, \qquad [4]$$

where U_b is the translation speed of the dune and Λ is the dune height. Porosity of well sorted sand ranges from 0.3 to 0.4. To determine the bedload transport by weight, the solid density is multiplied by the right hand side of equation 4. For sand, a solid density of 2650 kg/m^3 is typically used.

Simons and others (1965) found good agreement using this idealized view of the dunes as triangles with laboratory flume data. Willis and Kennedy (1975) stated that the assumption of triangular bedforms is sufficiently accurate for the computation of bedload transport.

To measure the bedform speed, Engel and Lau (1980) used the cross correlation of successive profiles to determine the lag in the bedform profile from one bathymetric measurement of the bedforms to the successive bathymetric measurement made at a later time. The cross-correlation method is explained in Nordin (1971).

Figure 2 contains successive days of bed profile data from the Mississippi River at St. Louis, Missouri, collected by the author and members of the St. Louis District Army Corps of Engineers (Corps) using the Corps bathymetric surveying boat, *M/V Simpson*. The bed profiles were collected in the same longitudinal transect one day apart. The bedforms, which are ubiquitous for this reach, are shown in figure 3.

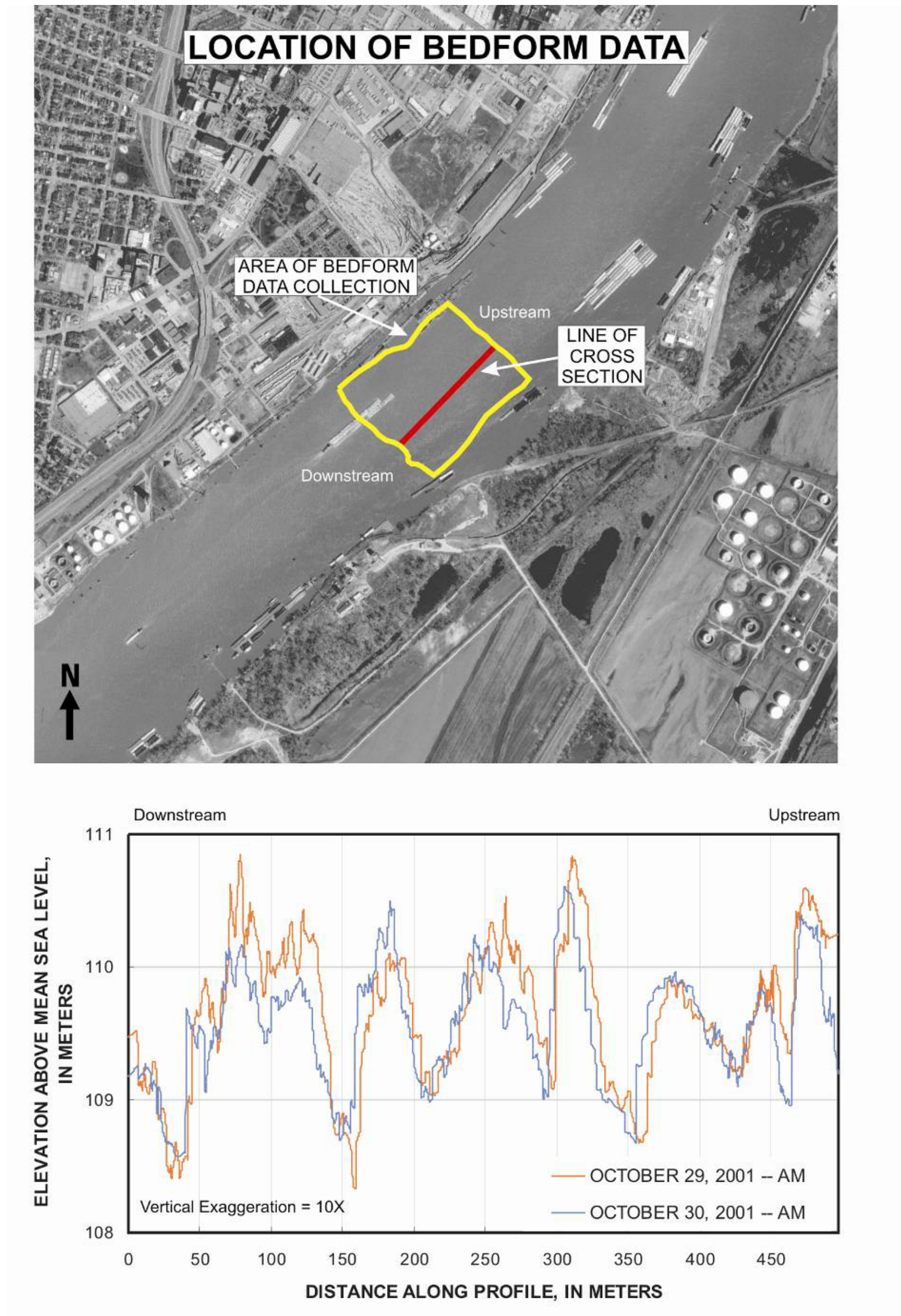

Figure 2. Streambed profiles for the Mississippi River at St. Louis, Missouri for successive days (October 29 and 30, 2001).

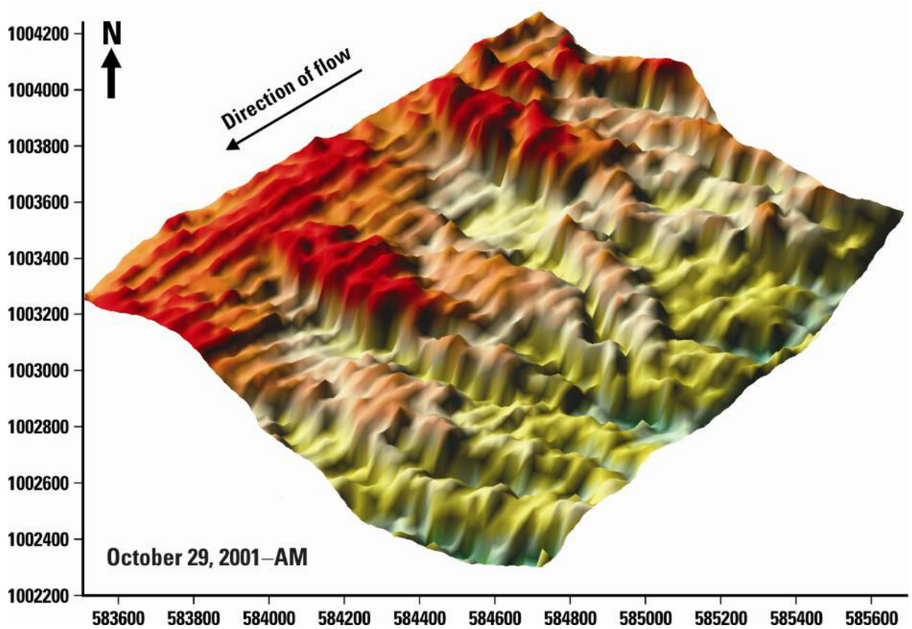

Figure 3. Channel bathymetry for Mississippi river at St. Louis, Missouri, for October 29, 2001.

Figure 4 shows the cross correlation between the two data sets when a lag distance is introduced in the data from October 29. The maximum correlation corresponds to the approximate distance the bedforms have moved, in this case 4.5 m, yielding a bedform speed, U_b= 4.5 m/day (5.2 x 10^{-5} m/s).

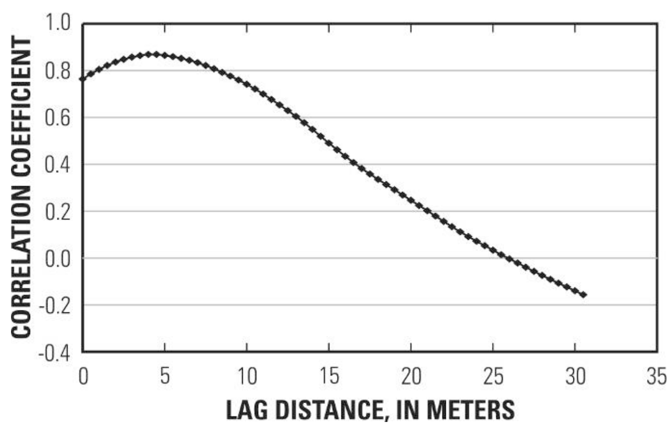

Figure 4. Lag distance versus correlation coefficient, Mississippi River at St. Louis, Missouri

Determining the exact bedform height can be a challenge and quite uncertain. Engel and Lau (1980) developed an equation that computes the unit bedload transport by weight, as

$$q_b = 1.32\gamma_s(1-\lambda_p)U_b\overline{\xi}\,, \qquad [5]$$

where γ_s is the unit weight of the sediment and $\overline{\xi}$ is the average value of ξ, the absolute value of the departure of bed elevation from the mean bed level $\overline{\eta}$, or $\xi = |\eta - \overline{\eta}|$. To determine the bedload transport by weight, the solid density is multiplied by the right hand side of equation 5, and for sand, a solid density of 2650 kg/m3 is used. Equation 5 has the benefit of not requiring a decision on bedform height, using a systematic computation of $\bar{\xi}$ to characterize the geometry of the sediment. Engel and Lau (1980) based part of the derivation of this equation on the assumed elevation of zero transport of sediment inferred from pressure distribution measurements made by Jonys (1973). The Mississippi River at St. Louis data for a flow rate of 4,050 m^3/s yield a value for $\bar{\xi}$ of 0.41 m. Using a porosity of 0.36, the bedload transport in this location of the reach was 4.08 t/m d (metric tons per meter per day).

The above methods take a two-dimensional approach to the bedform velocimetry method and require unique identification of a bedform as it progresses through time. Abraham and Kuhnle (2006) developed a methodology, ISSDOT (Integrated Section Surface Difference Over Time), that allows for the computation of the bedload at discrete gridded sections positioned laterally and longitudinally along the river bed. The method computes the sediment transport rate across the grid boundaries over time and has been tested in both the Mississippi River and in a controlled laboratory flume environment.

Bedload Measurement Using Virtual Velocity of the Bed-Material Sediment

The virtual velocity of the bed-material sediment was defined by Haschenburger and Church (1998) as "the total distance traveled (possibly incorporating multiple steps) by individual grains divided by the measurement interval." The volumetric bedload per unit width can be computed as

$$q_b = (1 - \lambda_p) V_b d_s , \quad [6]$$

where V_b is the virtual velocity of the bed-material sediment and d_s is the active depth of the streambed sediment movement. Once again, to determine the bedload transport by weight, the solid density is multiplied by the right hand side of equation 6 and for sand, a solid density of 2650 kg/m^3 is used.

Two methods have been used to determine the virtual velocity. Lagrangian techniques follow individual particles as they are transported, whereas Eulerian techniques examine particle velocities moving across a particular cross section of the channel.

Some of the earliest Lagrangian experiments were conducted by Hubbell and Sayre (1964) whereby they used radioactive sand released on the North Loup River near Purdum, Nebraska. Virtual velocity was determined by monitoring the magnitude of radioactivity along the channel in real time. Van Rijn (1984) describes the laboratory experiments of Fernandez Luque, who determined the virtual velocity by using high-speed photography to track individual particles in a laboratory flume. Haschenburger and Church (1998), using magnetically tagged stones as tracers, were able to install these tracers in the bed prior to a flow event and locate them after the event. Knowing the duration of flows sufficient to mobilize the bed sediments (mobilization time), they determined the virtual velocity for the sediments during the event as the distance traveled by the tracer divided by the mobilization time.

The advent of the acoustic Doppler current profiler (ADCP) has resulted in tremendous possibilities for measurement of flow properties. The impact of the ADCP on the density, speed, quality, and spatial resolution of water velocity data acquisition is unprecedented in riverine hydraulic research. The ADCP can measure volumetric flow rate by not only measuring the water velocities in three dimensions throughout the water column, but also by tracking the bed to determine the distance traveled as the ADCP is moved in a boat across a stream cross section. The streamflow discharge measurement accuracy of the ADCP is greatly diminished during streamflow conditions due to streambed mobilization. The inaccuracy is caused by the inability of the ADCP to determine the distance traveled because the bed moves relative to the ADCP movement, thereby introducing a bias. If the ADCP maintains a stationary position over a mobile bed, the bottom tracking feature of the ADCP indicates that the ADCP is moving (commonly known as "bottom-track bias"). In this mobile bed situation with a stationary ADCP, the indicated speed of ADCP movement should correspond to the virtual velocity of the bed-material sediments.

The ability of the ADCP to determine virtual velocities of bed-material sediments for potential bedload transport determination has been recognized for some time. A few investigators have made good progress in making measurements of virtual velocity using the ADCP (Rennie and others ,2002; Villard and others,2005). Rennie and others (2002) did an extensive field investigation of the capabilities of the ADCP to measure the virtual velocity on the gravel-bedded Fraser River, British Columbia, Canada. They investigated the sampling area insonified by the ADCP and the inherent error when multiple beams measure the bottom track, and, consequently, the virtual velocity, at multiple points along the bed. By taking a detailed look at the actual bottom track beam velocities, they determined that a large sample time, approximately 25 minutes, was needed to arrive at a stable average of the virtual velocity. Kenney (2006) took a simpler approach by using the "distance made good" report from the ADCP divided by the time of the observation to estimate the virtual velocity of the bed material.

The active depth of the streambed is dependent on streamflow discharge (Hollingshead, 1971; Slaymaker, 1972; Madej, 1984; and Haschenburger and Church, 1998). As such, active depth of the streambed is not constant and accuracy of the bedload transport rate in equation 6 is dependent on accurate measurement of both the active depth and the virtual velocity. Three methods to estimate the active depth of the streambed are discussed in this paper.

For streambeds with bedforms, Hubbell and Sayre (1964) present a method to estimate the active depth in a sand-bed stream by utilizing the streambed profile (figure 5). Multiple segments of a reach are determined by going from the upstream start of the surveyed section to the first trough downstream that is deeper than the starting trough. This method is repeated for the entire reach of surveyed channel. The active depth for each reach is estimated to be the difference between the mean bed elevation in the reach and the elevation of the upstream trough.

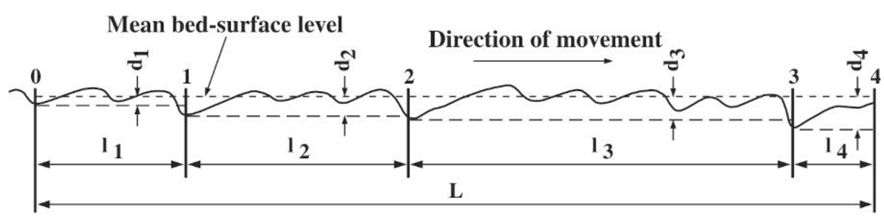

Figure 5. Schematic diagram illustrating method for estimating average depth of streambed sediment movement from Hubbell and Sayre (1964).

A second method that can be used to estimate the active depth is to assume a constant value of active depth by use of the classical definition of bedload being within two grain diameters of the bed (Einstein, 1950). Thus,

$$d_s = 2D_{50}, \qquad [7]$$

where D_{50} is the median diameter of the bed material.

The third method is to assume that active depth varies with particle diameter, D_*, and transport stage parameter, T, with equations developed by van Rijn (1984)

$$d_s = 0.3 D_*^{0.7} T^{0.5} D_{50}, \qquad [8]$$

$$D_* = D_{50}\left[\frac{(\rho_s/\rho - 1)g}{\nu^2}\right]^{1/3}, \qquad [9]$$

$$T = \frac{\left(u_*'\right)^2 - (u_{*,cr})^2}{(u_{*,cr})^2}, \qquad [10]$$

where ρ_s is the sediment density, ρ is the water density, g is the acceleration of gravity, υ is the kinematic viscosity, u_*' is the grain shear velocity, and $u_{*,cr}$ is the critical bed-shear velocity according to Shields (1936). Van Rijn (1984) computed the shear velocity as a product of the mean flow velocity, U, as

$$u_*' = \frac{g^{0.5}}{C'}U, \qquad [11]$$

Where C' is the Chezy coefficient related to the grain roughness defined as

$$C' = 18\log\left(\frac{12H}{3D_{90}}\right), \qquad [12]$$

and D_{90} is the size of the bed material where 90 percent of the bed material is finer.

Using these three methods for determination of the active depth, bedload transport rates using the bedform velocimetry and virtual velocity methods were determined for the Missouri River at St. Charles, Missouri, for a streamflow discharge of 3,160 m^3/s on June 19, 2002. Channel bathymetry was determined by hydrographic surveying, whereas virtual velocity of the bed-material sediments was determined at several locations longitudinally along a bedform by anchoring the boat at the various locations (figure 6). The bed profile for June 19, 2002 is not smooth because the sensitivity was set too high on the digital fathometer during the bathymetric survey for that day, however, the geometry and general dimensions of the bedforms are easily identified.

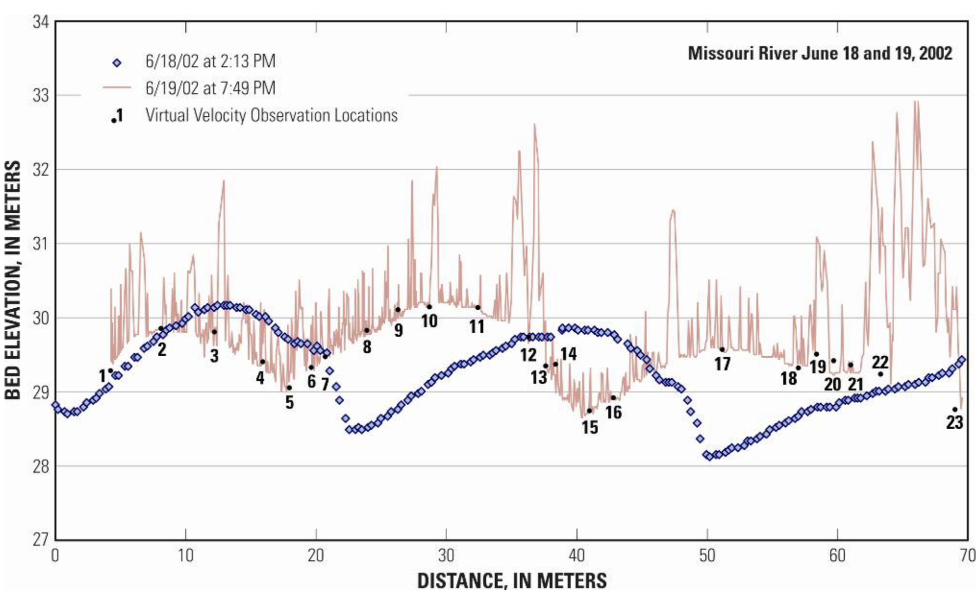

Figure 6. Streambed profiles for Missouri River at St. Charles, Missouri, for successive days (June 18 and 19, 2002).

Comparing the bed bathymetry for the previous day (June 18, 2002; discharge of 3,340 m^3/s) with the June 19, 2002data, indicates a bedform velocity of 0.523 m/hr. Using the measured bedform height of 1.25 m, the bedload transport rate of 13.3 t/m d was computed using the bedform velocimetry method. Figure 7 shows the bed bathymetry for June 18, 2002, with the red line delineating the longitudinal locations for the data in figure 6.

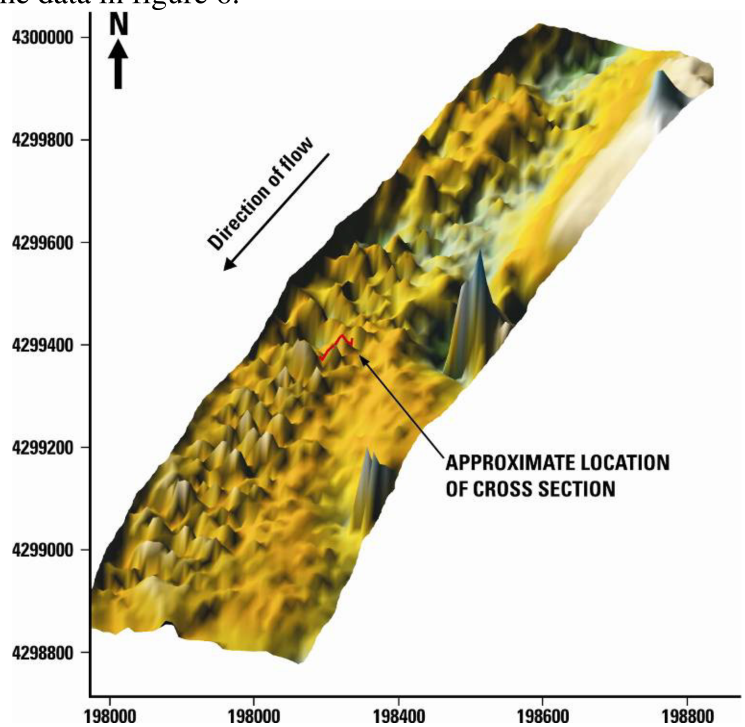

Figure 7. Channel bathymetry for Missouri River at St. Charles, Missouri, for June 18, 2002.

The virtual velocity at each location in figure 6 was determined as the ADCP "distance made good" divided by the elapsed time of the observation. Local observations of virtual velocity at each location are represented by the arrow sizes in figure 8. Table 2 contains the actual numerical values for the virtual velocities at each location. The larger values of virtual velocity are on the stoss side of the bedforms, with the largest of the three bedforms showing the largest virtual velocities, which is expected as the taller bedforms are more exposed to the flow. The spatially-averaged virtual velocity for the reach was determined to be 0.156 m/s.

Table 2. Virtual velocity for 23 longitudinally distributed locations for the Missouri River at St. Charles, Missouri on June 19, 2002.
[m, meter; s, second]

Location	Depth of Flow m	Virtual Velocity m/s
1	6.72	0.039
2	6.56	0.067
3	6.31	0.066
4	6.21	0.117
5	6.16	0.150
6	5.98	0.249
7	6.04	0.217
8	5.98	0.314
9	5.91	0.347
10	5.86	0.407
11	6.27	0.264
12	6.27	0.096
13	6.66	0.166
14	6.65	0.075
15	6.87	0.096
16	6.69	0.140
17	6.44	0.099
18	6.39	0.177
19	6.50	0.105
20	6.59	0.034
21	6.50	0.196
22	6.62	0.108
23	7.10	0.126

The active depth of the streambed was determined for each of the three methods previously mentioned. Assuming a porosity of 0.36 and a solid density of 2650 kg/m^3, the estimated active depths and bedload transport rates were computed and are presented in table 3.

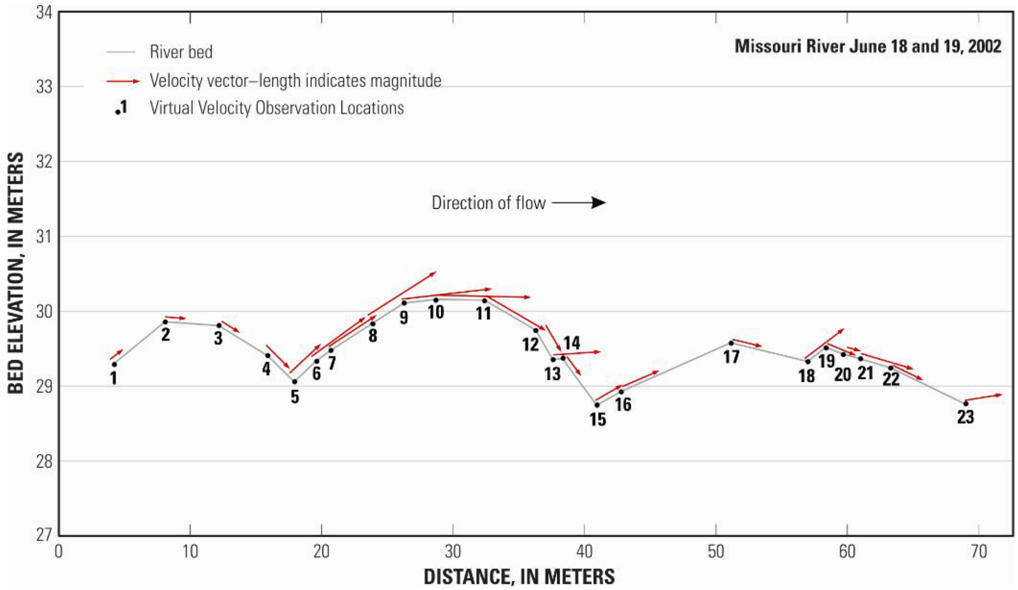

Figure 8. Virtual velocity at each location on the Missouri River on June 19, 2002.

Table 3. Bedload transport rates estimated for the Missouri River at St. Charles, Missouri for June 19, 2002.
[-- no value; m, meter; t, metric tons; d, day]

Method of Bedload Transport Rate Estimate	Active Depth of Streambed m	Bedload Transport Rate t/md
Bedform Velocimetry	--	13.3
Virtual Velocity; Hubbell and Sayre (1964) for active depth	0.61	13,900
Virtual Velocity; Einstein (1950) for active depth	0.00062	14.2
Virtual Velocity; van Rijn (1984) for active depth	0.0016	37.4

Disadvantages of Various Bedload Measurement Methods

Bedload has proven to be one of the most difficult fluvial processes to measure. Each of the three methods mentioned: physical traps, bedform velocimetry, and virtual velocity of bed-material sediments have limitations.

The use of a trap-type sampler has the advantage of directly collecting the samples and allowing the investigator to physically examine the bedload sediments as they are collected. In addition, outside of weighing the samples, little post-processing of the data is necessary to arrive at an estimate of bedload transport. However, there are numerous disadvantages to using a trap-type sampler in a sand-bed river. First, any direct sampling device placed on the bed disturbs the flow and rate of bedload transport, which is particularly problematic in sand-bed streams as entrainment of the sediments is more easily influenced by flow disturbance. The standard direct contact bedload sampler tends to dig into the bed when deployed (Rubin and others, 2001). The sampler must be placed squarely on the bed surface to properly sample. Bed irregularities, which are common in sand-bed streams with the various bedforms, can cause problems in regards to the fit of the sampler to the bed. Second, the mesh bag on the trap samplers can clog or allow small bed sediments to escape. Emmett (1980) states that bed sediments around the size of the mesh (0.20 – 0.25 mm) have this tendency. Third, the spatial and temporal variability of bedload transport require both a large number of samples and great deal of effort. Once the field samples are collected, additional laboratory time is required to analyze the samples. Lastly, placing a sampler at depth in flows capable of mobilizing bed-material sediments is difficult and dangerous.

Measurement of bedload transport using bedform velocimetry is fairly reliable provided that bedforms are prominent and the bathymetric measurements are of sufficient accuracy. For sand-bed systems in the transitional regime, where plane bed and/or ripples predominate, the bedform velocimetry method cannot be applied because no prominent bedform features are present to time their movement. The bedform velocimetry method requires collecting successive bathymetric measurements over time, typically requiring boat-mounted surveying capability, which limits the application to those locations that are accessible by boat. The requirement of a manned boat has been mitigated somewhat with the recent developments of both small remote-controlled boats equipped with acoustics and Global Positioning System (GPS) and the use of tethered boats with radio-telemetered data capabilities to relay the bathymetric information to shore. More problematic for the smaller rivers is the necessity of successive measurements and the resulting additional time required to collect the data.

Unlike large rivers, the streamflow discharge for small rivers often does not remain steady for sufficient length of time to provide a unique comparison of the streambed bathymetry for the same streamflow discharge. Traditional development of relations between flow characteristics and bedload transport often requires that the bedload transport be determined for a discrete flow condition. If the flow regime changes considerably during data collection, the bedload transport is an integration of the bedload transport lumped over the range of flow conditions, rather than an indication of the influence of a discrete flow on bedload transport. As such, a balance must be struck between allowing enough time between bathymetric surveys to have discernable bedform translation and the change in streamflow discharge.

Using the Lagrangian tracer method for virtual velocity in sand-bed rivers is likely not possible at this time (radioactive tracers are not an option). With the advent of the ADCP, the virtual velocity method shows great promise. ADCPs are becoming less expensive and can be deployed with relative ease from numerous stationary platforms (bridges or boats). Rennie and others (2002) found that for accurate determination of the mean of the virtual velocity of gravel sediments, data must be collected for approximately 25 min. Due to the spatial variability of bedload transport, multiple locations across the cross section will need to have the virtual velocity determined. As such, a 25- minute requirement per cross-section location is not practical for operational use, especially in small rivers where time of data collection is limited to determine relations between discrete flows and bedload transport. However, it is speculated that the length of time necessary for accurate determination of the virtual velocity in sand-bed sediments will be less than that of gravels, as gravel bedload transport is more episodic and unsteady when compared to sand bedload transport.

The most difficult task is likely arriving at a method to adequately estimate the active depth of the streambed sediment movement. The indirect method of Hubbell and Sayre (1964) requires a bathymetric survey and bedforms to be present. In the experiment on the Missouri River at St. Charles, Missouri, for June 19, 2002, the large inaccuracies of the Hubbell and Sayre method are apparent in the overestimate of the bedload transport. Although the estimate of bedload transport was close to the bedform velocimetry method (table 3), the use of a constant active depth of the streambed using the Einstein (1950) definition of bedload does not follow the findings of active depth dependence on streamflow (Hollingshead, 1971; Slaymaker, 1972; Madej, 1984; and Haschenburger and Church, 1998). The method of van Rijn (1984) provides an estimate of active depth that varies with flow regime and results in an estimate of bedload transport for the June 19, 2002, Missouri River at St Charles data that is reasonably close to that of the bedform velocimetry method (table 3).

Lastly, although the ADCP has shown utility in determining the virtual stream velocity, the velocity measured is that of the sediment in motion at the surface of the streambed. Just as the surface velocity of the water does not equal the mean velocity of the water column, the surface velocity of the streambed sediments is likely not equal to the mean velocity of the active depth of streambed sediment movement. Use of the surface velocity of the bed sediments to represent the virtual velocity for the entire active depth would result in an overestimate of the bedload transport rate.

Research Direction: Determination of the Active Depth of the Streambed

The virtual velocity method holds great promise. While development of the procedures and software necessary to fully utilize the ADCP in determination of the virtual velocity needs to continue, a focus of research should be directed to the development of sensors and/or methods to accurately determine the active depth of the streambed sediment movement.

Fluidization of the bed sediments results in inter-granular contact which in turn increases the porosity of the bed sediments in the area of fluidized layer. Accurate measurement of the porosity gradient in the streambed would provide the necessary information to determine the active depth of the streambed sediment movement. Possible technologies to explore for development of a sensor to measure the porosity gradient include geophysical electrical resistivity techniques and acoustic techniques such as that utilized in the Roxanne™ system (Any use of trade, product, or firm names is for descriptive purposes only and does not imply endorsement by the U.S. Government)..

Summary

Bedload transport in sand-bed rivers can be measured either through direct measurement with a physical trap, such as a Helley-Smith sampler, or through indirect measurements such as the bedform velocimetry and virtual velocity methods. Direct measurements are problematic in that a sampler placed directly on the streambed disturbs the flow and rate of bedload transport, which is particularly problematic in sand-bed streams as entrainment of the sediments is more easily influenced by flow disturbance.

Bedform velocimetry has been proven to be a good method, provided that prominent bedform features are present. The bedform velocimetry method works particularly well in large sand-bed river systems where the flow remains steady for sufficient time to allow successive bathymetric surveys at the same streamflow discharge. Small sand-bed rivers are often too unsteady to apply this method to arrive at a relation between the bedload transport and a discrete streamflow discharge. Bedload transport values of 4.08 and 13.3 metric tons/day/m were measured using the bedform velocimetry method at the Mississippi River at St. Louis, Missouri, on October 29-30, 2001, and the Missouri River at St. Charles, Missouri, on June 18-19, 2002, respectively.

Virtual velocities of the streambed sediments can be readily measured using the bottom-track bias inherent in ADCP instruments. Knowledge of the virtual velocity allows estimation of the bedload transport if the active depth of streambed sediment movement is known. A bedload transport rate of 13,900 t/m d for the Missouri River at St. Charles, Missouri, on June 19, 2002, was estimated by measuring the longitudinal averaged virtual velocity of the streambed sediment and estimating the active depth of the streambed sediment movement by using the Hubbell and Sayre (1964) method from the longitudinal streambed profile. More reasonable estimates of the bedload transport of 14.2 and 37.4 metric tons/day/m occur when the assumption of the classical definition of bedload depth of Einstein (1950) and the van Rijn (1984) equations to predict active depth were used, respectively. The virtual velocity method holds great promise, but the development of sensors and/or methods to accurately determine the active depth of the streambed sediment movement is needed to enable more confident estimates of bedload transport.

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