

SECTION 4. SEDIMENT TRANSPORT AND DEPOSITION

Cesium-137 Measures Erosion Rate and Sediment Movement--	
J. C. Lance, S. C. McIntyre, R. R. Lowrance, B. L. Campbell and R. L. Miller.....	4-1
Application of ¹⁰ Be to the Study of Erosion and Sediment Transport--	
Louis Brown, Fouad Tera, Nathalie Valette-Silver, Milan J. Pavich, Jeffrey Klein and Roy Middleton.....	4-10
Factors Affecting Accuracy of Bedload Sampling--	
David W. Hubbell and H. H. Stevens, Jr.....	4-20
Sediment Transport in Goodwin Creek--	
J. C. Willis, R. W. Darden and A. J. Bowie.....	4-30
A Laboratory Study of Sediment Transport under Ice--	
James L. Wuebben.....	4-40
Accuracy of Sediment Discharge Estimates--	
Thomas H. Yorke and Janice R. Ward.....	4-49
Local Scour at Drop Structures--	
Emmett M. Laursen, Matthew W. Flick and Brian E. Ehlers.....	4-60
Sediment Sources and Sediment-Transport Curves--	
K. M. Nolan, R. J. Janda and J. H. Galton.....	4-70
Missouri River Sediment Transport Relationship--	
Hsieh Wen Shen, Wen-Chiang Wang and Wayne G. Dorough.....	4-80
Denudation Rates in Selected Debris-Flow Basins--	
W. R. Osterkamp and J. E. Costa.....	4-91
Missouri River Aggradation and Degradation Trends--	
Warren J. Mellema and Tsong C. Wei.....	4-100
Patterns of Sediment Deposition in Shallow Bays--	
Flora Chu Wang, William H. McAnally, Jr. and J. S. Wei.....	4-110
Colorado River Sediment Transport in Grand Canyon--	
E. L. Pemberton and T. J. Randle.....	4-120
Probability Distributions for Bedload Transport--	
William P. Carey and David W. Hubbell.....	4-131
Bed Form Predictors for High Sediment Loads--	
Jeffrey B. Bradley and James D. Graham.....	4-141
Suspended-Sediment Transport, Lake Tahoe Basin--	
J. H. Galton and K. M. Nolan.....	4-152
Issues in Sediment Research in Rivers of the U.S.--	
Robert H. Meade and Randolph S. Parker.....	4-162

SECTION 4

SEDIMENT TRANSPORT AND DEPOSITION

CESIUM-137 MEASURES EROSION RATE AND SEDIMENT MOVEMENT

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ABSTRACT

Soils of the world are labeled with cesium-137 as a result of fallout from nuclear tests. The radioactive cesium moves with the clay because it is tightly adsorbed. Therefore measurements of the redistribution of radioactive cesium can be used to estimate erosion and sediment deposition rates. Measurements of cesium-137 concentrations in uneroded soils across the southern U.S. indicated that the cesium-137 input was proportional to the average annual rainfall. Vertical distribution of cesium-137 within the profile was related to soil properties rather than to rainfall. Cesium-137 concentrations decrease exponentially from the surface downward in uncultivated soils but concentrations are uniform throughout the plow layer of cultivated soils. Therefore, different techniques are needed to measure erosion rates on cultivated land, forest land and range land. Similar concentration profiles in a forest and an adjacent clear-cut area with a steep slope in Oklahoma indicated little erosion or mixing of the soil profile by the forest harvesting procedures. Comparisons of total cesium-137 activities in a forested area and a nearby peanut field in the coastal plain of Georgia indicated extensive losses of clay from the cultivated field. Thus, techniques for accurately estimating volumes of soil need additional refinement but the method shows promise as a useful tool for measuring soil loss and sediment deposition and for understanding erosional processes.

INTRODUCTION

Better methods of measuring erosion and deposition are needed to evaluate the loss of soil productivity due to erosion and downstream damages from sediments eroded from agricultural lands. Methods are needed to simultaneously measure erosion, deposition and sediment movement from the same field or plot. These kinds of data are urgently needed to verify non-point pollution, and soil erosion-soil productivity models.

Measuring soil loss from rainfall runoff plots is probably the method that has been used more than any other to assemble a soil loss data base. Also, a great deal of data has been gathered in watershed studies. The use of rainfall simulators on different plots with various rainfall intensities and durations is currently producing the greatest volume of erosion data. One disadvantage of these methods is that only net movement of sediment from a plot or watershed is measured. Soil that moves and is deposited without leaving the field is not measured. Measurements of Cesium-137 concentrations in the soil have the potential to measure sediment movement from fields and to measure the redistribution of soil within a field. However, these techniques need considerable development before they can be used to measure erosion and

deposition and track sediment movement through a watershed.

Soils of the world are labeled with cesium-137 as a result of fallout from nuclear tests. The radioactive cesium moves with the clay because it is tightly adsorbed. Therefore measurements of the redistribution of radioactive cesium can be used to estimate clay movement which can be related to erosion rates. Cesium concentrations decrease exponentially from the surface downward in uncultivated soils but concentrations are uniform throughout the plow layer of cultivated soils. Therefore, different techniques are needed to measure erosion rates on cultivated land, forest land and range land.

Ritchie et al (1974) correlated the fraction of cesium-137 loss from cultivated land with soil loss calculated by the Universal Soil Loss Equation (USLE) for several eastern watersheds. Then Ritchie and McHenry (1975) used that data along with data from studies using small runoff plots (Graham, 1963, Menzel, 1960 and Rogowski and Tamura, 1970) to show a logarithmic relationship between soil loss and cesium loss. The equation $x = 0.88 y^{1.18}$ best fit the data where soil loss was plotted vs. radionuclide loss as a percentage of radionuclide input. Ritchie and McHenry (1975) suggested that other equations probably would be needed for soils from other regions. Kachanoski and De Jong (1984) developed a model that predicts the amount of cesium-137 remaining in soil as a function of time and erosion rate. Their analysis showed that changes in the cesium-137 contents of Saskatchewan soils could be used to estimate erosion rates between 0.5 and 10 kg M⁻²y⁻¹ with reasonable precision if the sampling interval is at least 15 years.

The studies described in this report were initiated to refine methods for calculating soil loss from fields and sediment movement on fields, to extend the method to rangelands and forest land and to use the cesium-137 technique to develop a better understanding of erosion and deposition.

METHODS

Soil samples at 2-5 cm depth intervals have been collected from uneroded sites at locations across the southern U.S. from Arizona to Georgia to characterize the cesium input and movement in different soils and under different climatic conditions.

Samples were taken using a scraper plate technique that allows accurate calculation of the density of each layer of the soil. Gamma ray spectrometric analyses were made using a multi-channel analyzer and a lithium drifted crystal. Samples were dried at 105°C and 1 Kg subsamples were placed in Marinelli beakers. Cesium-137 was expressed on a mass basis (millibecquerels per gram dry weight) and on an area basis (becquerels per square meter). Several sites are being studied intensively and preliminary findings from two sites will be used as examples to illustrate the kind of work being done. Cesium-137 concentrations in a forest at the top of a 12% slope that had been clear-cut were compared with cesium-137 activities along the slope. This area in Southeastern Oklahoma was used to determine the effect of clear-cutting a pine forest on soil movement. Cesium-137 concentrations

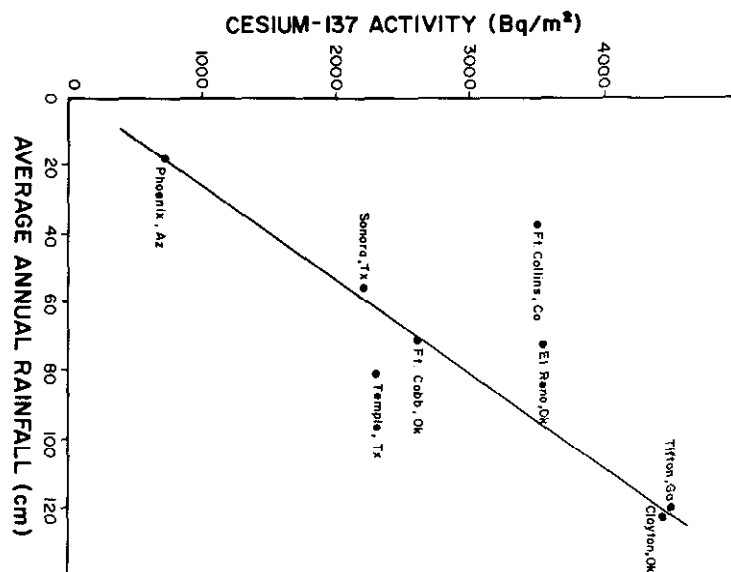


Figure 1. Total cesium-137 activities at different rainfall levels across the southern U.S.

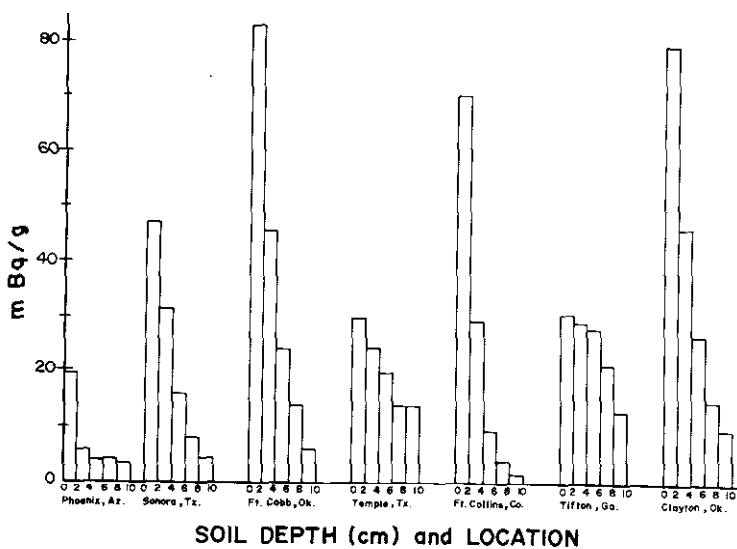


Figure 2. Cesium-137 concentration profiles in soils across the southern U.S. (Average annual rainfall increases from left to right see Figure 1).

in a peanut field and a nearby forested area in Georgia were compared to study soil movement in a cultivated Coastal Plain soil.

CESIUM-137 DISTRIBUTION ACROSS THE SOUTHERN U.S.

The total cesium-137 content of the soils collected at 8 locations across the southern U.S. appears to be proportional to the rainfall (Fig. 1). This is in agreement with results by Cox and Fankhauser (1984) on a much smaller scale for Hawaii. The cesium-137 activities ranged from 736 becquerels/meter² (Bq/m²) in the desert near Phoenix, AZ (19 cm/yr rainfall) to 4500 Bq/m² near Tifton, Georgia (120 cm/yr rainfall).

One site, near Fort Collins, CO had considerably more cesium-137 than would be expected for a relatively low rainfall zone (38 cm/yr). This may be due to its proximity to the Nevada test site. Although other factors such as wind patterns affect radionuclide deposition, it appears that rainfall data should be helpful in at least making approximate estimates of radionuclide inputs to soils.

The vertical distribution of cesium-137 within the soil profile did not appear to be dominated by rainfall amounts (fig. 2). For example, one of the highest rainfall sites, Clayton, OK (122 cm/yr) had high concentrations of cesium-137 near the surface and most of the activity was contained in the top 10 cm of soil. The Temple, TX site is in a lower rainfall zone (80 cm/yr) but concentrations were fairly evenly distributed through the top 10 cm with relatively low concentrations in the top 2 cm. The cesium-137 also penetrated farther into the soil at the Tifton site. It was the only site where less than 90% of the activity (73%) occurred in the top 10 cm of soil. Soil characteristics such as clay content and the type of clay probably determined the distribution of cesium within the profile. The high sand content of the soil at the Tifton site probably allowed penetration while cracking in the soil at the Temple site due to swelling clays allowed some penetration due to the migration of clay from the surface into the soil.

Tillage mixes the cesium fairly evenly through the plow layer in cultivated soils but the distribution of cesium in the profile must be considered in using cesium-137 to measure erosion on range or forest soils. More research will be needed to develop the techniques or models for estimating soil loss from these soils using cesium-137.

SEDIMENT MOVEMENT FOLLOWING FOREST CLEARCUTTING

Samples were collected along a transect down a steep forested slope (12%) in Southeastern Oklahoma that had been clear cut and replanted 3 years earlier and from an undisturbed forested area at the top of the slope. The slope length was about 300 m and the soil was a Carnasaw Silt Loam. A scraper plate technique was used to excavate 30.5 x 45.7 cm pits in 2 cm thick layers. Cesium-137 concentrations declined exponentially from the surface to the 10 cm depth with little cesium-137 present below the 10 cm depth (fig. 3). These well defined concentration profiles were found at all positions on the slope indicating that the logging operation had not mixed the cesium-137 from the surface

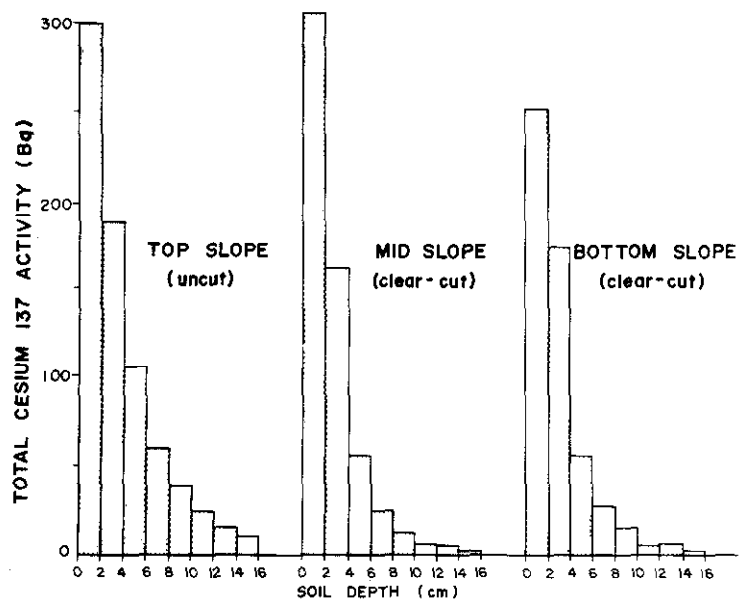


Figure 3. Cesium-137 concentration profiles at different slope positions following clearcutting (Forest remained at the top of the slope).

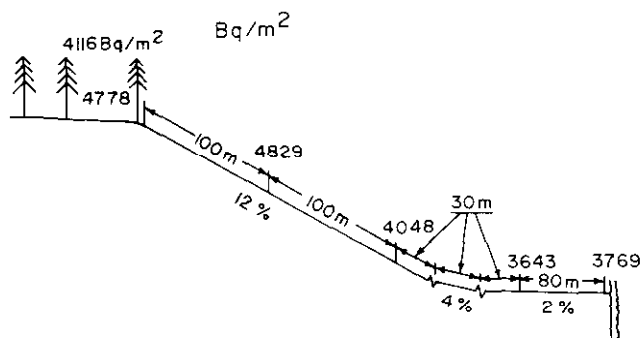


Figure 4. Total cesium-137 activities at different slope positions following clearcutting (Forest remained at the top of the slope).

into deeper layers. This suggests only limited disturbance of the soil due to logging operations. It also indicates the slope was only slightly eroded and that little deposition of sediment occurred at any of the sampling sites.

The sampling site 100 m from the top of the slope indicated no loss of cesium-137 (fig. 4). The site at 200 m indicated possible slight loss and sites at 290 and 370 m downslope indicated a definite loss. Its somewhat surprising that the sites showing the most cesium loss were at the foot of the slope where the land was sloping slightly (about 2%). Apparently the runoff flow concentrated as it moved down this long, steep slope resulting in soil loss at the base of the slope rather than deposition. The low erosion rates probably are due at least in part to one management practice. When the land was replanted to trees the soil was deep plowed or "ripped" to a depth of about 45 cm at 3 M intervals down the slope and pine seedlings were planted in the furrows to increase their survival rate. These deep furrows which were placed across the slope roughly on contour undoubtedly reduced runoff and collected some sediment.

The sampling sites at the base of the slope indicated a loss of about 17% of the cesium-137. However its difficult to estimate soil loss from the measurement. If we use the formula of Ritchie and McHenry (1975) the loss would be 24.6 mt/ha. However, that formula was developed primarily from cultivated soil where the cesium-137 has been mixed throughout the plow layer. The cesium-137 loss at this site originated from the surface layer which contains about 4 times the cesium concentration that would be found in the surface layer if the same total input of radionuclide has been mixed throughout the plow layer by cultivation. Therefore the soil loss at the base of the slope may be only 1/4 the 24.6 mt/ha value or about 6 mt/ha. Then the average soil loss for the entire slope probably would be less than half that value. Although the accuracy of such calculations could be questioned, they support the earlier inference that soil loss from this slope was quite low and would still support that inference even if large errors were assumed. More samples are needed from this slope to account for spacial variability and to measure cesium-137 contents at slope breaks to check for deposition. These could be core samples to measure total cesium-137 content at various points on the slope because the concentration profile with depth has been fairly well established by the earlier samples. These rough estimates indicate that the sensitivity of the cesium method for these soils may be greater than that estimated for Saskatchewan soils by Kachanoski and De Jong (1984).

SEDIMENT MOVEMENT ON A COASTAL PLAIN SOIL

Samples were collected along a transect down a 6% slope using the scraper plate technique in a peanut field located in the Coastal Plain region near Tifton, Georgia. The soil was a Tifton sandy loam. The cesium-137 was mixed fairly uniformly throughout the upper 24 cm of the soil profile. The cesium-137 input was measured by sampling two sites in nearby pine forests where the trees were at least 30 years old. The sampling site at the top of the slope in the peanut field showed little loss of cesium-137 but all of the sites down the slope showed

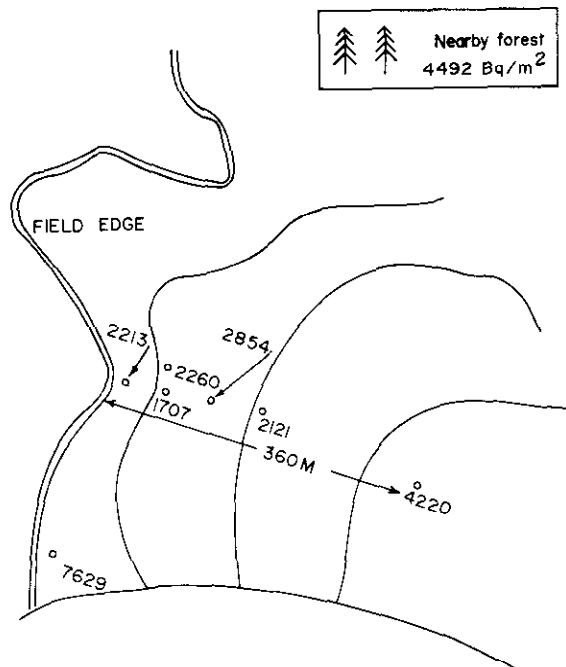


Figure 5. Total cesium-137 activities at different slope positions in a peanut field and in a nearby forest.

considerable loss except one site in the corner of the field that showed an accumulation of cesium-137 (fig. 5). The cesium-137 losses ranged from about 36 to 62% of the input with an average loss of 50%. A 50% loss of cesium-137 indicates a soil loss of 88 mt/ha has using the equation of Ritchie and McHenry (1975).

This soil loss would be equivalent to the loss of approximately 1 cm of top soil. This would not seem to be a particularly severe loss during a cultivation period of 30 years (since the radioactive fallout began). However, a 50% loss of cesium-137 must represent a 50% loss of the clay fraction which adsorbs the cesium-137. This is probably a fine clay fraction and may indicate a substantial loss of colloidal material. Its reasonable to assume that such a loss could severely reduce soil productivity even though the estimated soil loss does not seem to be extremely high. More research is needed on the effect of these losses on soil productivity. Its possible that the cesium-137 loss would be a better indication of loss in soil productivity than measurements of gross soil loss.

The area of deposition in the corner of the field was about 5% of the total field area. Although the cesium-137 activity in the small area of deposition was almost twice that in uneroded sites (fig. 5), that cesium-137 accumulation would represent only a fraction of the cesium-137 lost from the entire field. Again the cesium-137 probably was adsorbed to fine clay fractions which were not deposited on the field. The wetland area adjacent to the field will be sampled in an attempt to trace the movement and final destination of the cesium-137. Also, measurement of all of the sediment deposited at the field edge and in the wetland may indicate a higher erosion rate than that calculated by the formula of Ritchie and McHenry (1975).

These results show that techniques for accurately estimating values of soil loss using cesium-137 movement need additional refinement but the method shown promise as a useful tool for measuring soil loss and sediment deposition on upland soils. The method also shows promise as a way to better understand the erosion process by studying the movement of the fine clay fractions containing cesium-137.

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APPLICATION OF ^{10}Be TO THE STUDY OF EROSION AND SEDIMENT TRANSPORT

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ABSTRACT

The radioactive isotope ^{10}Be (1.5×10^6 yr) is produced in the atmosphere by cosmic rays and descends on the earth primarily in rain and snow. Present evidence indicates that it is generally tightly bound to soil, having soil to water partition coefficients of 10^5 to 10^6 , thereby requiring periods of 10^5 yr to penetrate a few meters of fine grained regolith. Measurements of the dissolved isotope in ground and river water indicate that to a good approximation the isotope's subsequent transport is by particulate matter. Based on this characteristic we have introduced a new method for examining the erosion and transport of soils and sediments. By measuring the ^{10}Be concentration in the sediment carried by a stream from a basin, one can compare (providing sediment transport data are available) the rate at which atoms leave the basin with the rate at which they are deposited in it. Owing to the large inventory known to exist in soils and to decay there, this ratio must be less than unity for steady state ^{10}Be deposition and erosion. Measurements are presented and their possible utility as diagnostics discussed.

INTRODUCTION

Suspended-sediment sediment yield is determined by measuring the suspended-sediment concentration of a stream at a gauging station. From these data average values of the annual load are calculated, which provide an average loss per unit area of the basin. This procedure, so simple in concept, entails difficulties of which the inaccuracies of the individual measurements are only a part. For many basins sediment transport occurs primarily in easily recognizable high magnitude events generated by rains and snowmelts. The distribution of the magnitude of these events gives rise to large uncertainties of the annual sediment load requiring years of observation to obtain a reliable average. The expense of this approach obviously limits its applicability, but the accumulation of much less frequent measurements of sediments taken sporadically for a variety of flow conditions allows one to determine the two parameters p and j of the sediment rate equation (Graf, 1971)

$$G = pQ^j \quad (1)$$

where G is the daily suspended sediment transport in tons (2000 lbs) and Q is the stream flow in ft^3/sec . Once this empirical equation has been determined annual sediment loads can be "synthesized" by using the much more easily obtainable stream flow data. The same parameters, p and j , may be derived from the daily measurements, allowing a third estimate of sediment load to be calculated.

^{10}Be APPLIED TO SEDIMENT TRANSPORT

This approach complements the traditional methods of evaluating sediment yield. It depends on the assumption that the ^{10}Be transported from the basin is carried overwhelmingly in the particulate matter. Evidence, in addition to what has already been discussed, for the validity of this assumption comes from two experiments, one on elemental Be dissolved in river water, the other on ^{10}Be dissolved in river water. Measures and Edmonds (1982) have determined the concentrations of dissolved elemental Be in the Amazon, Orinoco and Yangtze Rivers. Milliman and Meade (1983) report the sediment loads of these rivers. If one accepts the Be concentration to be 3 ppm, an almost canonical value for soils and rocks (Wedepohl, 1969), then one can compare the two modes of Be transport. All three rivers have less than 2% carried in solution. Similarly, a measurement of dissolved ^{10}Be in Potomac River water at Point of Rocks, Maryland (Valette-Silver et al., 1985) is consistent with a sediment to water partition coefficient of 10^5 and accounts for less than 5% of the ^{10}Be carried in the sediment load of that river.

These examples indicate that the ^{10}Be carried from a basin is transported overwhelmingly in the sediment load. Thus by knowing the sediment load and the ^{10}Be concentration in the sediment one knows how many atoms of the isotope are transported from the basin per year, a number that can be compared with the deposition rate on the basin. This ratio, I_B , is called the ^{10}Be basin index. It is defined as

$$I_B = nM/qA \quad (2)$$

where M is the annual sediment load in grams, n , the ^{10}Be concentration in atom/g, q the deposition rate in atoms $\text{cm}^{-2} \text{yr}^{-1}$, and A the basin area in cm^2 .

We have examined the ^{10}Be content of a wide variety of soils and find generally large inventories. With so much of the isotope locked up in soil where a large fraction of it must ultimately decay, the basin index, I_B , must be less than 1 for steady state conditions. A value greater than 1 means more ^{10}Be is leaving the basin than is incident on it and infers that erosion is in a transient condition in which inventories built up in soils over long periods are being eroded at accelerated rates.

The weakest assumption of the basin inventory method is that concentrations of ^{10}Be measured in the river or estuarine sediments are representative of the sediment discharge, i.e., the weighted mean concentration of ^{10}Be in the sediment transported by the stream. Our approach is to obtain samples from a few separate locations, e.g., bottom, flood plain or bank, without bias towards a particular grain size, and see how consistent they are. This is hardly satisfactory but allows us to gain experience from which proper sampling techniques may evolve.

We have examined two different kinds of basins: (1) basins defined by U.S.G.S. Water Resources Division stations for streams draining generally similar kinds of terrain and (2) continental sized basins with world wide distribution.

Given the difficulty and cost of these measurements, any new quantitative approach should be helpful. We have obtained data using the isotope ^{10}Be in a technique that complements the traditional practice and also provides information about the recent history of erosion from a basin.

INTERACTION OF ^{10}Be WITH SOIL, SEDIMENT AND WATER

Beryllium-10 is produced in the atmosphere by cosmic rays striking nitrogen and oxygen with subsequent deposition onto the earth in rain and snow; dry deposition, which accounts for less than 10% in regions having 100 cm of annual rainfall, may be important in extreme deserts. The deposition rate is difficult to determine and is presently uncertain by possibly 25%.

Measurements made over the past four years in collaboration with G. Stensland of the Illinois State Water Survey, D. Bogen of the Environmental Measurements Laboratory, New York City, and A. Yoshinaga of the Mauna Loa Observatory lead us to a tentative rate for temperate latitudes with 100 cm/yr of precipitation of $1.3 \times 10^6 \text{ atom cm}^{-2} \text{ yr}^{-1}$, a value in agreement with recent work by Monaghan et al. (1985).

Prior to 1977 measurements of ^{10}Be in nature were sorely restricted because of its long half life. This is best illustrated by noting that from 10^9 atoms only one will normally decay in a day. It had been observed where concentrations exceeded 10^9 atom/g : deep ocean sediments, manganese nodules, meteorites and lunar rocks and soil. The introduction of mass spectrometry with a nuclear particle accelerator has changed this situation dramatically, as it allows measurements of concentrations of the isotope in rocks and soils as low as 10^5 atom/g to be made routinely. For a survey of the instrumental techniques see Litherland (1980) and of the applications see Brown (1984).

Once deposited ^{10}Be adheres to the components of soil with a high tenacity; soil to water partition coefficients for Be^{+2} in excess of 10^5 being observed near neutral pH (Nyffeler et al., 1984). This implies that once landed the isotope's subsequent lateral movement is with particulate matter while its vertical movement into the soil column is largely in solution, a process consistent with the high partition coefficient. The measurement of ^{10}Be as a function of depth in a core taken 16 m into the Virginia Piedmont shows the isotope to be in the top few meters. The difference between the ^{10}Be found in the profile and what might be expected from total retention can be accounted for as leaving through erosion of the surface (Pavich et al., 1985). Direct measurements of dissolved ^{10}Be in various natural waters are just beginning (Valette-Silver et al., 1985). At present we see no evidence that dissolved ^{10}Be in surface runoff is an important loss for the soil profile. Our working hypothesis is that ^{10}Be escapes the site inventory only by radioactive decay or particle transport through erosion. Vertical transport within the column does not alter the inventory. If this picture is true for a wide range of geological conditions, the utility of ^{10}Be as a tracer of sediments is assured. At present the assumption of very high retention under a wide variety of conditions is not proved, but there is evidence for its validity where tests have been made and no contradictions yet observed except in exceptionally acidic conditions.

Water Resources Division Stations

Through the co-operation of the Water Resources Division we have obtained sediment samples from the vicinities of a number of stations for which transport data exist. We have measured the ^{10}Be concentrations for 28 of these basins. Table 1 lists for them the station number and name, the basin area, the station altitude and co-ordinates, together with sediment transport data calculated from data obtained from WATSTORE. Values for the annual sediment load are extracted from the WATSTORE data by means described in the caption. Note that serious discrepancies exist between measured-load and the D-load, even though both rely on the same set of measurements. For determining the average value, M , given in the last column and used for calculating the erosion index we use a weighed average given by

$$M = (w_1 M_m + w_2 M_d + w_3 M_q) / (w_1 + w_2 + w_3) \quad (3)$$
$$w_1 = N_d, w_2 = 0.1 N_d N_f / (N_d + N_f), w_3 = 5 N_q N_f / (N_q + N_f)$$

where N_d , N_q and N_f are the number of daily, sporadic and flow days respectively and M_m is the measured load, M_d the D-load and M_q the Q-load. This weighting is arbitrary. The greatest weight is assigned the daily measurements, the least to D-load calculations and an intermediate amount to Q-load calculations. Note for example: the Pee Dee River at Pee Dee has mixed data. The weights calculated are $w_1 = 2064$, $w_2 = 183$ and $w_3 = 264$. The preferences given Q-load calculations result from their being taken at representative flow conditions.

The ^{10}Be data are reported in Table 2 that divides the 28 basins according to the location of the stations and the basins: the Coastal Plain, the Piedmont and the Highlands. The annual rainfall, used to calculate the deposition rate q , is taken from Climate Atlas of the United States.

1) The Coastal Plain. All six of these basins have erosion indices less than 1 with none greater than 0.5. Two basins with extraordinary low indices, the Contentnea and the Edisto, had samples of well sorted, uncoated quartz sand giving low ^{10}Be concentrations. Excluding these two gives an average for the index of $0.25 \pm 68\%$. Although consistently less than 1 these values do not necessarily infer steady state.

2) The Piedmont. These 16 stations, located between the Fall Line and the mountains are more varied in their indices. Nine have values greater than 1 and only one of the remaining five has an index less than 0.5, and this station is located 5 km below the Roanoke Rapids Dam, the farthest downstream of three dams forming large reservoirs. These dams reduce the sediment yield more than a factor of 10 below any of the other basins of the group. Excluding this station gives an average index of $1.79 \pm 69\%$, a value 7 times as great as that of the Coastal Plain. This disparity of sediment transport between the Coastal Plain and the Piedmont has been noted before by Meade (1969, 1982) and Trimble 1974, 1975, 1977). A destructive agriculture practiced, especially in the last century, in the Piedmont set enormous quantities of soil in motion, little of which has reached the Coastal Plain. Indeed Meade (1969) has estimated that "modern sediment loads are probably 4 to 5 times greater than they would be if the area had remained undisturbed by man."

TABLE

Station No.	Station Name	Area sq. km	Alt. m.	D— days	Q— days	F— days	meas. ld. k — ton/yr	D — load k — ton/yr	Q — load k — ton/yr	mean ld. k — ton/yr
1 01470500	Schuylkill R at Berne	919	95	11978	0	13307	80.81	30.67	0.00	78.30
2 01481000	Brandywine Cr at Chadds Ford	743	46	5430	49	23069	52.39	126.90	122.00	60.53
3 01481500	Brandywine Cr at Wilmington	813	21	11597	0	13908	60.28	115.90	0.00	63.16
4 01540500	Susquehanna R at Danville	29060	131	913	102	29046	2120.00	1708.00	1145.00	1767.63
5 01545600	Young Womans Cr nr Renovo	120	238	364	225	6883	1.92	1.90	1.89	1.90
6 01553500	Susquehanna R, W Branch	17734	130	730	101	16437	617.00	476.00	329.00	498.39
7 01567000	Juniata at Newport	8687	111	11942	0	30908	273.00	419.00	0.00	282.82
8 01570500	Susquehanna R at Harrisburg	62400	88	5410	0	34162	3220.00	2950.00	0.00	3198.54
9 01638500	Potomac R at Point of Rocks	25996	61	8374	0	32383	711.00	1203.00	0.00	747.21
10 01643000	Monocacy R at Frederick	2116	71	3104	0	20102	125.00	424.00	0.00	148.84
11 01664000	Rappahannock R at Remington	1606	77	11297	0	14975	100.70	454.50	0.00	119.78
12 02035000	James R at Cartersville	16206	50	220	218	21975	68.00	594.00	550.00	470.29
13 02075500	Dan R at Paces	6604	98	9070	0	12427	700.00	861.10	0.00	708.80
14 02080500	Roanoke at Roanoke Rapids	21720	13	0	92	10630	0.00	0.00	71.30	71.30
15 02084160	Chicod Cr at Road 1760	117	2	2511	0	9144	3.43	2.66	0.00	3.37
16 02087500	Neuse R nr Clayton	2978	39	0	105	20881	0.00	0.00	174.00	174.00
17 02089500	Neuse R at Kinston	6972	3	0	117	3932	0.00	0.00	83.00	83.00
18 02091500	Contentnea Cr at Hookerton	1888	5	0	61	20393	0.00	0.00	17.60	17.60
19 02091970	Creeping Swamp nr Vanceboro	70	5	0	128	7773	0.00	0.00	0.76	0.76
20 02098198	Haw R at Moncure	4375	47	0	43	2216	0.00	0.00	149.00	149.00
21 02100500	Deep R at Ramseur	904	128	0	47	6458	0.00	0.00	60.00	60.00
22 02105769	Cape Fear R at Lock 1	13610	-1	0	114	5557	0.00	0.00	253.00	253.00
23 02131000	Pee Dee R at Pee Dee	22870	8	2064	53	16464	546.00	564.00	277.00	519.02
24 02151500	Broad R at Boiling Springs	2266	195	0	32	11366	0.00	0.00	492.00	492.00
25 02175000	Edisto R nr Givhans	7070	6	2027	94	16392	26.10	38.40	9.37	24.01
26 02198500	Savannah R nr Clio	25510	4	0	102	18113	0.00	0.00	172.00	172.00
27 03176500	New R at Glen Lyn	9759	454	0	31	20833	0.00	0.00	90.50	90.50
28 03453500	French Broad R at Marshall	3450	502	0	89	15342	0.00	0.00	482.00	482.00

A listing of the Water Resources Division stations for basins of the eastern United States that were used in this study. The three columns headed D-, Q- and F-days give the number of days for which certain kinds of data are available: D-days have daily measurements of suspended sediment and stream flow; Q-days have suspended sediment and stream flow measured sporadically (QW data), generally to match varying flow conditions; F-days have daily measurements of stream flow. The next four columns give calculations of the annual sediment load, all in units of 10^3 ton (2000 lbs) per year. The "meas. ld." is the average yearly total of daily measurements and as such is the closest to a measurement in the direct sense. The next two columns give annual loads calculated from Eq. (1) but with the parameters p and j determined from different sets of data: the "D-load" uses p and j determined from daily measurements and thus uses these data a second time; the "Q-load" uses p and j determined from the sporadic measurements (QW data). The "mean. ld." is calculated from Eq. (3).

TABLE 2

Station Name	Conc. 10 ⁹ g ⁻¹	%	Rain cm/yr	Yield mg/cm ²	Index	Notes
THE COASTAL PLAIN						
Chicod Cr at Road 1760	4 0.309	16	128.2	2.6	0.49	
Neuse R at Kinston	4 0.193	4	127.0	1.1	0.13	2
Contentnea Cr at Hookerton	3 0.093	5	111.8	0.8	0.05	1, 2
Creeping Swamp nr Vanceboro	4 0.400	44	128.2	1.0	0.24	2
Cape Fear R at Lock 1	4 0.120	9	117.4	1.7	0.13	2
Edisto R nr Givhans	4 0.141	21	113.8	0.3	0.03	1
THE PIEDMONT						
Schuylkill R at Berne	4 0.411	39	116.1	7.7	2.10	
Brandywine Cr at Chadds Ford	4 0.131	8	110.5	7.4	0.67	3
Brandywine Cr at Wilmington	3 0.546	53	112.9	7.0	2.62	3
Susquehanna R at Harrisburg	4 0.311	23	100.9	4.7	1.10	
Potomac R at Point of Rocks	3 0.421	4	109.4	2.6	0.77	
Monocacy R at Frederick	2 0.850	4	111.4	6.4	3.75	
Rappahannock R at Remington	3 0.726	1	105.1	6.8	3.60	
James R at Catersville	3 0.320	19	109.7	2.6	0.59	
Dan R at Paces	3 0.276	67	111.8	9.7	1.85	
Roanoke at Roanoke Rapids	4 0.117	13	114.3	0.3	0.02	1, 2, 9
Neuse R nr Clayton	4 0.257	18	128.2	5.3	0.82	2
Haw R at Moncure	2 1.046	4	123.7	3.1	2.01	2, 4
Deep R at Ramseur	2 0.476	67	123.7	6.0	1.78	2
Pee Dee R at Pee Dee	4 0.636	13	115.6	2.1	0.87	5
Broad R at Boiling Springs	2 0.341	25	129.5	19.7	3.99	2
Savannah R nr Clio	4 0.696	28	120.4	0.6	0.27	2, 6
THE HIGHLANDS						
Susquehanna R at Danville	4 0.117	10	102.7	5.5	0.48	
Young Womans Cr nr Renovo	4 0.340	86	103.4	1.4	0.36	7
Susquehanna R, W Branch	4 0.601	20	102.7	2.5	1.15	
Juniata at Newport	4 0.164	8	102.7	3.0	0.36	
New R at Glen Lyn	4 0.725	25	101.6	0.8	0.46	8
French Broad R at Marshall	2 0.403	9	132.1	12.7	2.97	2

The integer number preceding the concentration is the number of samples for which the ¹⁰Be concentration has been determined. The number following the concentration is the single measurement standard derivation of the average.

Notes: (1) Very sandy sample.

(2) No daily sediment load data.

(3) The discrepancy between these two stations obviously lies in the ¹⁰Be data. The Chadds Ford sample was very sandy.

(4) Station located 91 m downstream from the Everett Jordan Dam. Reservoir began filling in September 1981 and the sediment data precede this.

(5) Reservoirs upstream. The first on the Pee Dee is Blewett Falls Lake located approximately 110 km from the station; the first on Black Creek, a tributary joining above the station, is about 95 km from the station.

(6) Reservoirs upstream. The first, Clark Hill Lake, is located approximately 277 km from the station.

(7) Drains Sproul State Forest.

(8) Dam about 70 km upstream.

(9) Station located about 5 km below Roanoke Rapids dam.

There are reservoirs on the Savannah, the closest about 277 km upstream from the U.S.G.S. station, and on the Pee Dee and its tributary Black Creek, their closest about 110 and 95 km, respectively, upstream from the station. Given the uncertainties of sediment transport time constants it is difficult to say whether these dams affect the erosion index. The Savannah River index would not stand out from the indices of the Coastal Plain stations. A study of sediment transport by the Savannah River at Clyo (Meade, 1976) discloses no change in sediment load since these reservoirs began to fill.

3) The Highlands. This group of six basins is too small and too diverse to allow us to make a generalization but is nonetheless instructive. The first four are from neighboring mountainous regions of Pennsylvania. Their indices are low as one might expect from regions of good forest cover. The lowest index applies to Young Woman's Creek, which drains the Sproul State Forest. Similar conditions apply for the New River station on the Virginia-West Virginia border. The high index for the French Broad River fits better with the Piedmont, and indeed land use in this basin has many similarities with the Piedmont.

Continental-sized Basins

For comparison with the Water Resources Division Stations we have included in Table 3 data from eleven basins that offer a limited, world-wide coverage, seven of them of continental sized areas. Sediment load data are from Milliman and Meade (1983) for all but the Narbada and the Tapti Rivers, which are from Borole *et al.*, (1982), and the rainfall data are from Rudloff (1981). The indices calculated from these data must be compared to those of the eastern United States with caution because the sediment load and rainfall data are less uniform. Furthermore, there is reason to doubt that

^{10}Be deposition has the same proportionality to rainfall in the tropics, where no measurements have been made, as in temperate climates because local rain cannot supply more ^{10}Be than is produced in the troposphere and transferred from the stratosphere. In such regions we have taken the maximum deposition rate to be $1.5 \times 10^6 \text{ cm}^{-2} \text{ yr}^{-1}$, a value that would indicate 115 cm/yr rainfall. Despite these limitations these basins are instructive.

The four Asian rivers have remarkably high indices. The Huang Ho would seem to be the world champion, both for its $140 \text{ mg cm}^{-2} \text{ yr}^{-1}$ yield and its index of 20. Note that it would be much higher were it not for the unusually low ^{10}Be concentration of $70 \times 10^6 \text{ atom/g}$, which was determined from very fine silty, not sandy samples. The explanation of this low concentration probably lies in the river eroding thick deposits of loess. A low ^{10}Be concentration is consistent with the origin of this loess being the accumulation of material ground up by the glaciers from bedrock and deposited there by wind. Such material would have had no initial concentration of the isotope and have had insufficient time since the last ice age to accumulate concentrations typical of mature soils. Whatever the geology may be, the erosion index of 20 demonstrates that the erosion of this basin is not only of huge rate but is transient. The Yangtze has similarities with the Huang Ho but draws on a smaller fraction of the basin made up of loess. The two neighboring Indian rivers, the Narbada and the Tapti, draw on very similar terrains. The difference in the ^{10}Be indices

(and the sediment yields) results from a dam on the Tapti that has reduced the sediment load.

TABLE 3

River	Area (10^{15} cm^2)	Sed. load (10^{12} g/yr)	Rainfall (cm/yr)	^{10}Be concn. (10^9 g^{-1})	Sed. yield ($\text{mg cm}^{-2}\text{yr}^{-1}$)	Index
ASIA						
Huang Ho	7.7	1080	37	3 0.070 78%	140	20
Yangtze	18.3	478	115	5 0.290 38%	26.1	5.1
Narbada	0.89	50	110	6 0.173 29%	56.2	6.8
Tapti	0.62	8	110	4 0.205 5%	12.9	1.9
SOUTH AMERICA						
Amazon	71.7	1200	115	3 0.212 17%	12.6	1.7
S. Francisco	6.4	6	100	1 0.136	0.9	0.1
Orinoco	9.9	225	115	8 0.035 26%	21.2	0.5
NORTH AMERICA						
Mississippi	32.2	210	80	10 0.362 26%	6.5	2.3
Columbia	6.7	8	50	1 0.378	1.2	0.7
Hudson	0.20	1.0	100	1 0.289	5.0	1.1
EUROPE						
Po	0.70	15.	82	1 0.120	21.4	2.4

Basins of world wide distribution. The rainfall for tropical regions is taken as 115 cm for calculating the erosion index for reasons given in the text. The three columns under " ^{10}Be concn." give the number of datum points, the concentration in atoms/g and the single measurement standard deviation of the average concentration in percent.

Two of the South American rivers have low indices, indicating steady state erosion. Both erode basins with a large fraction of mountainous terrain that has not had intemperate land use. The Amazon basin with its huge sediment load can hardly admit of an easy interpretation. Sediments from the weathering of the Andes, which account for 90% of the total (Meade *et al.*, 1985), mix with soils from tropical rain forests. We have three measurements of ^{10}Be concentrations in the Peruvian headwaters that average $47 \times 10^6 \text{ atom/g} \pm 64\%$, the kind of low value that we have come to expect from mountainous erosion. A similar situation applies to the Po, which erodes an Alpine terrain, and has an index of 2.4 despite the extremely high sediment yield. It would appear that high concentration valley sediments are mixed with mountainous material of low ^{10}Be concentration.

The Mississippi has a relatively high value, at least compared with the Coastal Plain stations of the eastern United States. The Columbia and Hudson rivers, both having only a single ^{10}Be datum, do not appear much different from other American highlands.

CONCLUSIONS

Beryllium-10 has chemical characteristics that make it useful in examining erosion quantitatively. There is direct and indirect evidence that to a good approximation ^{10}Be is retained by soil, and its subsequent transport is with the sediment. The method is being tested on a set of watersheds that are as nearly uniform in geology and climate as is possible and for which sediment transport data exist. Its principal weakness, recognized as such in advance, is the difficulty of selecting samples of sediment that match the average sediment transported by the stream from the basin. No criteria have yet been formulated for the acceptability of samples, and the samples measured for the 28 basins in the Water Resources group have been those collected by field hydrologists, who received only a brief description of what was desired. Nevertheless, this set of samples has given a fairly consistent picture, all things considered. Of particular note is the clear indication of enhanced sediment movement in the Piedmont but not in the Coastal Plain, in agreement with Meade's and Trimble's observations. Asian rivers show very high indices. This manifestation of accelerated erosion is in agreement with much other evidence that erosion is taking place at a higher rate than at earlier times, possibly since the introduction of agriculture on a wide scale but perhaps since the end of the ice age or even before, as a result of accelerated tectonic uplift.

The uncertainties of the indices, which should be reduced as sample selection guides are better worked out, must be compared with a range of values that spans two orders of magnitude. Three uses of the basin transport method present themselves: using ^{10}Be measurements to estimate sediment yield where conventional techniques are lacking, to help evaluate sediment yield where conventional data are discordant and to correct denudation rates from transient to steady state values.

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FACTORS AFFECTING ACCURACY OF BEDLOAD SAMPLING

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ABSTRACT

The accuracy of bedload measurements is affected by sampler calibration and sampling procedures. Both of these factors are greatly affected by the extreme temporal rate variations that occur naturally, even during steady-flow conditions, as a part of the process of bedload transport. Calibration curves for several commonly used samplers have been developed recently using a procedure that is intended to provide accurate calibration relations, despite the extensive rate variations that occur during calibration and normal sampler use. Analyses of sample sets selected from simulated transport-rate records show that maximum probable errors are large whenever temporal rate variations are not accounted for by repetitive sampling (multiple traverses), even when spatial sampling is fairly concentrated. Information on bed elevations during sampling may aid in improving the accuracy of sampled rates; however, an initial attempt to use such information was unsuccessful.

INTRODUCTION

Bedload-transport rates usually vary markedly with time in a cyclic manner at all points of transport in a stream cross section, even during steady-flow conditions. When the bed is formed into dunes, transport rates are zero, or almost zero, in the troughs of the dunes and a maximum at the crests. Although documentation is sparse, cyclic-rate variations also seem to occur even when bed forms are not pronounced. Transport rates measured over a 1.3-ft width at the center of a bedload trap (weigh pan 4) during two different laboratory flume runs with uniform 6.5-mm bed material illustrate typical patterns (fig. 1). Measurements made with basket samplers (Ehrenberger, 1931; Nesper, 1937) document comparable variations in several coarse-bedded European streams.

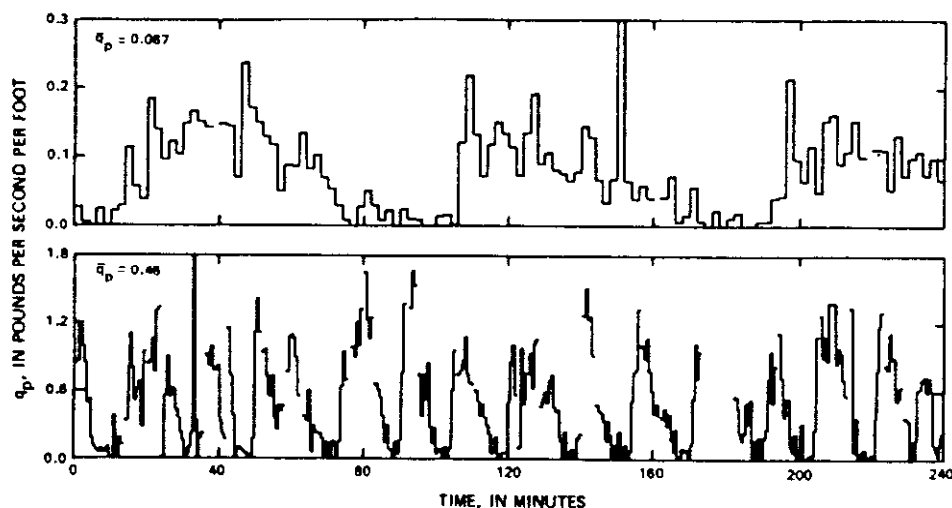


Figure 1.--Temporal records of bedload discharge in a laboratory flume with 6.5-mm bed material (from Hubbell and others, 1985).

When nondimensional rates, $q/\bar{q}$, in which q is any instantaneous rate and $\bar{q}$ is the long-term mean rate, are plotted as cumulative relative-frequency distributions, the graphs generally closely follow the probability distribution function, $F[q/\bar{q}]$, derived by Hamamori (1962) for dunes with superposed ripples:

$$F[q/\bar{q}] = 1/4 q/\bar{q} [1 + \ln 4 \bar{q}/q] . \quad (1)$$

Bedload-transport rates of gravel measured by Einstein (1937) and of sand sampled by Jobson (Helley and Smith, 1971) in the laboratory are plotted with Ehrenberger's (1931) field measurements and 6.5-mm data (trap) in figure 2 for comparison with Hamamori's (1962) distribution. Distributions from other sources show comparable agreement. It is evident from the distribution curves that bedload-transport rates vary from zero to approximately four times the mean as a natural consequence of the bedload-transport process in streams having bed material composed of sand and gravel.

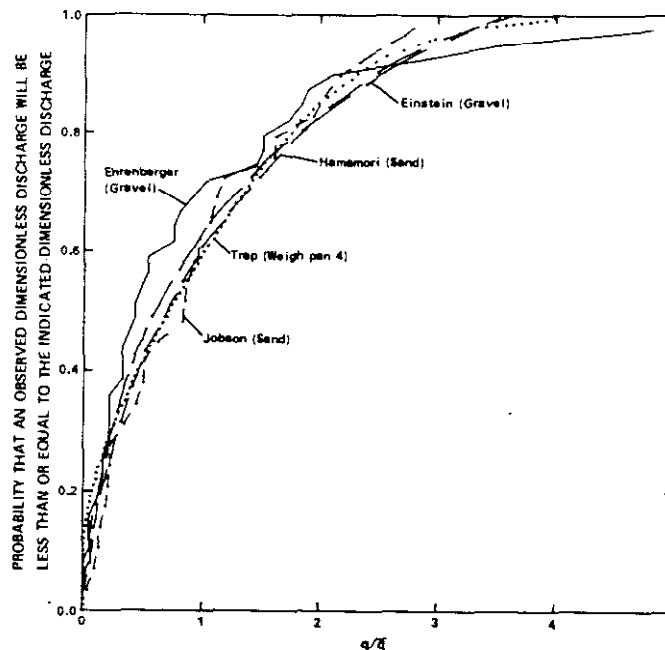


Figure 2.--Probability distributions of dimensionless bedload discharges for various bed materials (from Hubbell and others, 1985).

The persistence of extreme temporal rate variations, even at constant-flow conditions, substantially complicates the calibration of portable direct-measuring bedload samplers that rest on the streambed and can significantly affect calibration results. The variations also directly affect the sampling procedures that are required to provide reasonably accurate mean transport rates with such samplers. This paper briefly discusses the effects of rate variations on sampler calibrations; examines, for several different sampling procedures, the magnitude of measurement errors that could result with various hypothetical spatial distributions of mean rates due to natural temporal rate variations; and suggests how bed-elevation data might be used to minimize sampling requirements.

SAMPLER CALIBRATION

To provide accurate bedload discharges, samplers traditionally have been calibrated to define their sampling efficiency, E_s . Efficiency is defined as the ratio of the mass of bedload collected during any single sampling time, q_s , to the true mass of bedload that would have passed through the sampler's entrance width during the same time if the sampler had been absent, q_t , (Hubbell, 1964). In most calibrations, the efficiency has been determined as the ratio of the averages of a group of sampled rates and a group of corresponding measured (true) rates; efficiencies are defined for a range of steady-flow conditions. If, for practical purposes, the efficiency of a sampler is a constant value, then the extreme rate variations that inherently occur during the calibration procedure do not affect the determination of efficiency. However, if the efficiency of a sampler varies with the transport rate, then, because of rate variations, averages of the sampled and measured rates cannot be used to define efficiency without some degree of error (deVries, 1973).

Recent Calibration Experiments

Laboratory calibration of several variations of the Helley-Smith bedload sampler (Helley and Smith, 1971; Emmett, 1980), the VUV sampler (Novak, 1957), and the Arnhem sampler (Waterloopkundig Laboratorium, 1949) was recently completed by the U.S. Geological Survey in cooperation with the Federal Inter-Agency Sedimentation Project of the Federal Inter-Agency Advisory Committee on Water Data. The calibration program was conducted at the University of Minnesota's St. Anthony Falls Hydraulic Laboratory, Minneapolis, Minnesota, in a calibration facility large enough to accommodate prototype samplers and unscaled sediments. The facility consisted principally of an automated sediment trap beneath a slot in the floor of the channel which measured bedload-transport rates every 6 seconds. Four different bed materials were used during the calibrations. Sampling was conducted, during a series of runs, upstream from the trap at locations where the configuration of the streambed simulated natural field conditions. An initial assumption during the calibration program was that sampling efficiencies might vary with transport rate; hence, because of naturally occurring rate variations, average sampled and measured rates should not be used to define efficiencies. As a result, an alternative analytical procedure had to be developed.

Definition of Calibration Curves

The alternative calibration procedure is called the "probability-matching" method (Hubbell and others, 1985). It is based on the concepts that the distribution of a large group of widely varying sampled rates obtained at a particular location is a distorted form of the distribution of actual (true) rates at that location, and that the function required to adjust the distribution of sampled rates to agree with the distribution of true rates (the calibration curve) can be determined by comparing the distributions. The procedure involves separately ordering all q_s and q_t rates for a given run, from lowest to highest, and then plotting them as cumulative relative-frequency distributions. Values of q_s and q_t , selected at common probability (relative-frequency) levels from the distribution functions, then are plotted against each other to define the calibration curve. The procedure is believed to be valid if, among other things, the two distributions are comparably

ordered; that is, if the lowest sampled rate corresponds to the lowest true rate, and if the second lowest sampled rate corresponds to the second lowest true rate, etc.

The various runs with different bed materials provided information for defining calibration curves for each sampler for the 1.4- to 4.0-mm, 4.0- to 8.0-mm, and 11.3- to 32.0-mm size ranges. The curves shown in figure 3A and 3B for the standard 3- by 3-in sampler (no. 1), and sampler no. 3 [3- by 3-in entrance and 1.4 expansion ratio] illustrate the general character of the calibration results. The curve for a perfect sampler would plot as a single straight line that passed through the origin with a 1:1 slope. Because most calibration curves had some degree of nonlinearity, sampled rates are adjusted to true rates through use of the curves rather than by being divided by a constant efficiency.

BEDLOAD-SAMPLING PROCEDURES

The inherent temporal variations in bedload-transport rates that occur continuously at every position in a cross section seriously affect the accuracy of bedload measurements. If the mean transport rate at all points is uniform across a channel, a reasonable estimate of the cross section mean can be computed by averaging sampled rates defined throughout time at any single fixed location, or by averaging sampled rates obtained at a series of lateral locations. With both procedures, assuming overall sampling times are equivalent, the same rate population is being sampled. However, if the mean transport rate varies laterally, as it does at most measurement cross sections, then the temporal rate population at every point is different, and combining individual sampled rates from different lateral locations will yield a value of unknown accuracy. To quantify the approximate magnitude of sampling errors that could result from various sampling situations, a bedload-transport simulation model based on Hamamori's (1962) distribution has been developed. The model has been used to generate hypothetical temporal records of transport rates for various assumed lateral distributions of mean rates. These records, in turn, have been "sampled" by different procedures to withdraw sample sets comparable to those that might be obtained in real sampling situations; sampler efficiency is considered to be a constant value of 100 percent (a perfect sampler).

Simulation Model

The model produces simulated records of transport rates at individual points for selected values of mean transport rate, maximum bed-form height, and bed-form height-length ratio. Any or all of these variables can be changed systematically or randomly from one rate cycle to the next to create a varied record. For the sampling simulations described in this report, transport-rate records at different lateral positions in the cross section were generated using preselected mean transport rates ($\bar{q}$), a constant bed-form height-length ratio of 0.04 and either a constant maximum bed-form height of 0.6 ft, or a height that varied randomly about 0.6 ft. Typical simulated records, based on constant and variable maximum bed-form heights, are shown in figures 4A and 4B.

The lateral distribution of mean transport rates in many cross sections can be characterized in a general way with a quasi two-element representation. By assuming that a maximum mean rate, q_m , exists over a part of the stream width, W_m , while a minimum mean rate, q_n , exists over the remainder of the width, W_n ,

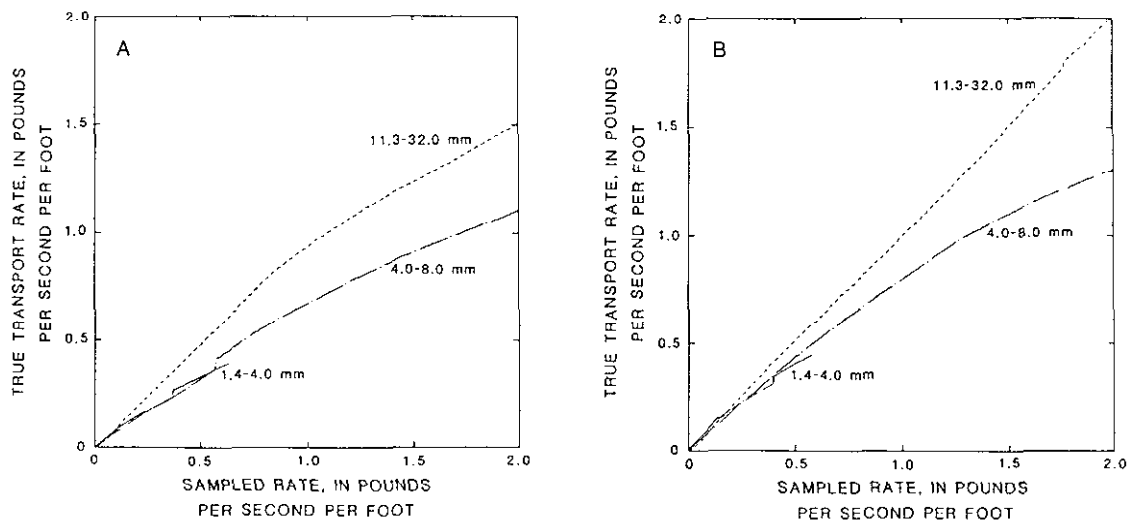


Figure 3.--Calibration curves for adjusting sampled rates of various size fractions of material collected with (A) the standard Helley-Smith bedload sampler (no. 1); and (B) Helley-Smith sampler no. 3.

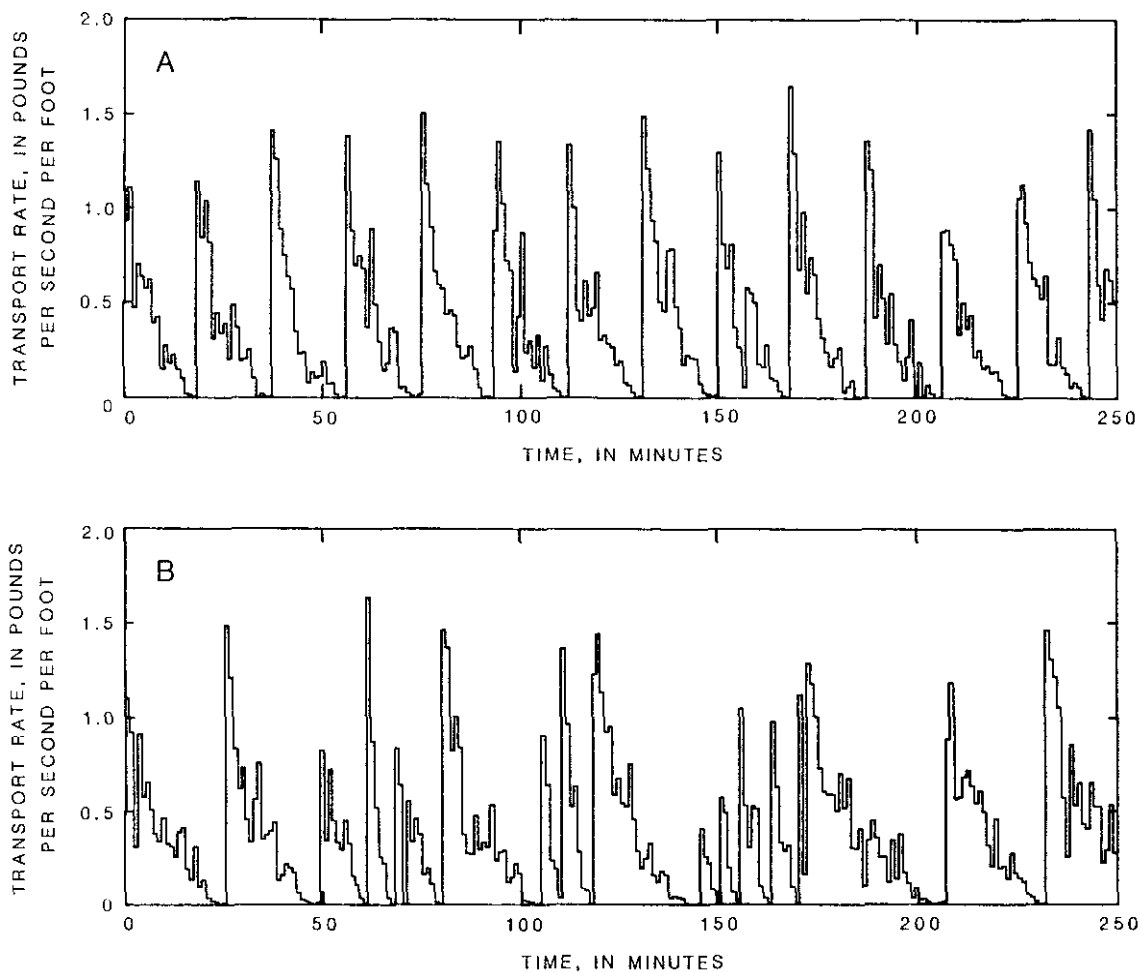


Figure 4.--Temporal records of bedload-transport rates generated with the simulation model, using (A) a constant maximum bed-form height; and (B) a randomly variable maximum bed-form height.

a variety of definable lateral distributions of mean transport rate can be created from selected q_m/q_n and W_m/W_n ratios. The lateral distributions can be defined systematically with even more detail if (1) q_m is subdivided into two rates, each of which is considered to occupy one-half of W_m , such that their average equals q_m and one rate is twice the other [$q = (4/3)q_m$ and $q = (2/3)q_m$] and (2) q_n is divided and apportioned over W_n similarly.

In an effort to examine maximum probable errors that might result from different sampling conditions, combinations of W_m/W_n and q_m/q_n were used to produce various lateral distributions of mean transport rates. In each distribution, two pairs of mean transport rates exist; however, the proportion of the width occupied by each pair, and the magnitude of the difference between the mean rates within each pair, varies from one distribution to another. For each distribution, the transport simulation model was used to generate four temporal records of transport; the records are based on $\bar{q} = (4/3)q_m$, $\bar{q} = (2/3)q_m$, $\bar{q} = (4/3)q_n$, and $\bar{q} = (2/3)q_n$. Each record is assigned to the appropriate lateral sampling positions, the locations of which are determined by the number of positions to be sampled and the W_m/W_n value. Sampling positions are spaced a fixed interval apart, and the first and last positions are one-half of the interval from the banks. The records, then, are "sampled," as might be done in the field, by initially selecting a rate from the record assigned to the first position nearest to the bank, and then progressively selecting rates from records assigned to each adjacent lateral sampling position at times that are a fixed interval (3 minutes) apart. When sampling in a cross section is completed, the process is repeated until the desired number of traverses, K , has been made. Individual "sampled" rates from each traverse then are averaged to provide a mean cross-section rate, and, in turn, sequential cross-section rates are averaged to produce final mean rates for different numbers of traverses. The entire process is repeated 50 times ($n = 50$), and variances of the 50 final rates for the different numbers of traverses are computed.

Variances of transport rates associated with cross-section samplings consisting of 1, 2, 4, and 20 sequential traverses at 4 and 20 lateral positions for combinations of W_m/W_n of 0.25, 1.0, and 4.0, and q_m/q_n values of 1.5, 3.0, 7.0, and 25.0 were determined. The $\bar{q}$'s used in the simulation model to generate records for each width element in the cross section were computed from values of q_m and q_n determined from the W_m/W_n and q_m/q_n ratios for each assumed distribution, and a constant cross-section mean transport rate of 0.4 lb/s/ft. Finally, the maximum probable percentage error, MPE, associated with each sampling situation was determined from:

$$MPE = (Q - \mu)100/\mu = (t \sqrt{s^2}) 100/\mu \quad (2)$$

in which Q is the mean rate computed with cross section rates from K traverses, μ is the true mean cross section rate, s^2 is the variance of the n mean rates, and t is the value from the t -distribution for the 99-percent confidence level and $n-1$ degrees of freedom.

Potential Sampling Errors

The most striking aspect of the results of the computations is how large the error can be when only a few traverses are made. In all simulated situations, sampling at 20 lateral positions with a given number of traverses yields better accuracy than sampling at 4 positions with the same number of traverses; however, as the number of traverses is increased, the magnitude of

potential errors for both procedures decreases markedly. In addition, for any given number of traverses with either 20- or 4-position traverses, when W_m/W_n is less than about 0.25, the maximum probable error increases as q_m/q_n increases; whereas, when W_m/W_n is greater than about 0.25, the effect of different q_m/q_n ratios is mixed and is not particularly significant.

Maximum probable errors associated with different numbers of traverses while sampling at 20 and 4 lateral positions from 2 distinctly different lateral distributions are plotted in figures 5A and 5B. The lateral distribution associated with each set of error curves also is diagrammed in the graphs. Although only one specific distribution is illustrated in each graph, the error curves are valid for any lateral arrangement of the subsection rates.

The lateral distribution of mean transport rates in figure 5A is moderately nonuniform ($W_m=W_n$, $q_m=1.5q_n$). The distribution in figure 5B is extremely nonuniform ($W_m=W_n/4$, $q_m=25q_n$). With the moderately nonuniform distribution, in order to reduce potential errors below about 30 percent, at least 4 traverses need to be made with the 20-position traverse; whereas, with the 4-position procedure, 10 to 20 traverses are required. Apparently, with this kind of distribution, when verticals are equally spaced, similar numbers of total samples provide comparable potential errors with either procedure. With the extremely nonuniform distribution, potential errors of less than about 30 percent can be attained only by sampling repeatedly for 9 or 10 traverses with the 20-position procedure and for about 100 traverses with the 4-position procedure.

The need for many repetitive samplings at each lateral position (multiple traverses) in cross sections where the spatial variations in mean rates are large seems to result because the inherent temporal rate variations are not accounted for by the spatial samplings. On the other hand, in cross sections where the lateral distribution of mean rates is fairly uniform, temporal rate variations apparently are represented to some degree by spatial samplings.

In these examples, the mean cross-section rate was determined by averaging the mean rates from all traverses included in the sampling sequence. This procedure is equivalent to determining the mean cross-section rate by averaging all samples collected at each lateral sampling position during the traverses, and then weighting each position average by the same proportion of the width. In general, overall accuracy is attained by accurately determining the mean rate at a sufficient number of individual positions to define the true lateral distribution of mean rates, and then weighting the individual mean rates by the proportion of the width they represent. Thus, in cross sections where the lateral distribution of rates is known approximately, it may be possible to reduce sampling effort by determining accurate mean rates, through repetitive sampling, at only those positions that are required to define the lateral distribution reasonably well, and then obtaining a properly weighted average rate.

Use Of Bed-Elevation Data

Because of rapidly changing transport conditions, or other similar restrictions, sampling at a series of verticals to obtain a number of replicate traverses may not be possible. Use of continuous records of bed-surface elevations at the sampling points, obtained with sonic sounders (fathometers) during sampling, offer a possible means for decreasing errors when only a few samples can be collected. By using the simple-wave equation and a bedload-continuity equation, together with some generalizing assumptions, it is possible to derive the relation, $q/q'' = y/y''$, in which q

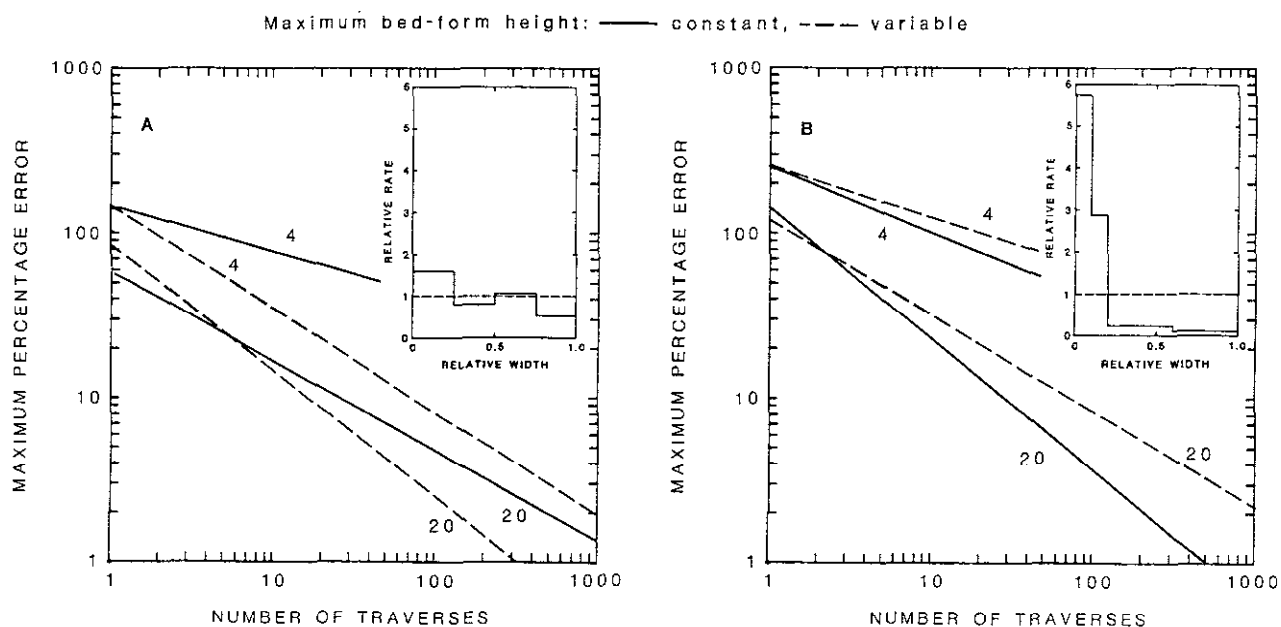


Figure 5.--Variation in maximum probable errors with number of sampling traverses for 20- and 4-position traverses at cross sections with different lateral distributions of mean transport rates (99-percent confidence level).

is the instantaneous bedload discharge at a point on the bed where the elevation is a distance, y , above the elevation of the base of the closest upstream trough, in which $q = 0$; and q'' and y'' are the largest discharge and maximum trough-to-crest distance, respectively, observed during the passage of a series of bed forms (Hubbell and others, 1985). Although this relation is not precise, it demonstrates that the transport rate at any location on the bed can be expected to correlate with the relative elevation, y/y'' . Thus, it is reasonable to assume that knowledge of the relative (or absolute) bed elevation at the time and place of a sampling can be used to adjust the instantaneous sampled rate to more accurately represent the mean rate at the sampling position. The following exploratory procedure for using bed-elevation information is based on this assumption.

True bedload-transport rates and corresponding relative bed elevations, which had been obtained for constant-flow conditions during a series of 15 laboratory flume runs with different bed materials, were arranged separately to form a pair of cumulative relative-frequency distributions for each run. Then, values of q and y/y'' were selected at common probability (frequency) levels from each pair of distributions to define a relation between these two variables. This procedure is meaningful provided the distributions are similarly ordered (the smallest q corresponds with the smallest y/y'' , the second smallest q corresponds with the second smallest y/y'' , etc.), and the error associated with each variable is small. The defined relations for the 15 runs produced a family of curves having, in general, a common functional form. By dividing transport rates from each curve by the mean rate for the run, the curves were normalized to the extent that they could be represented reasonably well by a single nondimensional curve ($q/\bar{q}$ as a function of y/y''). Because the dimensionless curve was developed from data for different bed materials

and flow conditions, it represents, to some degree, a general relation that possibly can be used to estimate the mean transport rate that corresponds to a given pair of measured q and y/y'' values.

Sampled rates obtained with sampler no. 3 during two runs with a bed-material mixture and corresponding measured relative bed elevations have been used to test the procedure. One hundred sets of three sampled rates that had been collected consecutively were selected randomly from the group of sampled rates that had been obtained during each run. Each rate was divided by its corresponding relative bed elevation to form a $q/(y/y'')$ ratio, and the three ratios were compared for consistency. Any $q-y/y''$ pair whose ratio differed by more than 50 percent from the next closest ratio was eliminated from the set. The remaining $q-y/y''$ pairs then were used to obtain a mean rate from the dimensionless curve, and the several mean rates were averaged to provide the best estimate of the true mean rate from that set of samples. Variances for each of the 2 test runs were computed using deviations of the 100 estimated mean rates from the known mean rate for the run. These variances, in turn, were used with the t-distribution (99-percent confidence level) to estimate the maximum probable percentage deviation (error) for means determined in this way. Results from the 2 runs indicated that the maximum probable errors were 179 and 236 percent. Similar analyses, in which the average of the 3 sampled rates from each of the 100 sets for a run was used to represent the estimated mean rate, gave maximum probable errors of 118 and 186 percent. Thus, the analyses show that using bed-elevation data in this way to enhance accuracy actually has the opposite effect, and that a simple average of rates from three consecutive samples can be expected, on the average, to yield a better estimate of the true mean rate than the average of rates adjusted by using the dimensionless curve. Although this particular application of bed-elevation information was not fruitful, other procedures may provide desired results.

SUMMARY

Laboratory and field measurements for constant-flow conditions demonstrate that bedload-transport rates of sand and gravel vary markedly with time, from zero, or almost zero, to about four times the mean transport rate. The variations in rates generally correspond with the probability distribution function derived by Hamamori (1962). The existence of these extreme temporal variations in transport rate complicate bedload-sampler calibrations and bedload-sampling procedures.

Calibration curves developed for a group of tested bedload samplers show that, generally, the relationship between true and sampled rates is not linear for particles ranging in size from 1.4 to 32.0 mm. The curves were defined by an analytical procedure based on comparisons of the distributions of measured (true) and sampled rates, rather than a comparison of average rates. The procedure is intended to provide accurate calibration relations despite the widely varying transport rates that occur naturally in streams and during calibration tests.

A model based on Hamamori's distribution has been developed to generate temporal records of bedload transport. Simulated records produced with the model have been used to examine errors that might result from using various sampling procedures for computing the mean transport rate in cross sections having different hypothetical lateral transport-rate distributions. Simulations show that (1) maximum probable percentage errors are extremely large whenever only a few samples are collected at each lateral sampling position; (2) for a given number of sampling repetitions, maximum probable

errors are smaller when 20 lateral positions, rather than 4, are sampled; (3) errors are greatest in cross sections where the lateral distribution of mean transport rate is extremely nonuniform; and (4) possibly, sampling effort can be decreased without increasing errors by sampling repetitively at only a few critically selected lateral positions. Based on theoretical considerations, it may be possible to use information on bed elevations at the times and places of sampling to enhance the accuracy of sampled rates. However, a preliminary scheme using relative bed elevations was unsuccessful.

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SEDIMENT TRANSPORT IN GOODWIN CREEK

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ABSTRACT

Concentrations of sands and fine sediments that have been collected since about 1980 in Goodwin Creek, MS, are used to develop tentative regression equations with depth of flow in the critical flow structures. Depth-integrated samples define the best estimates of the average sand concentrations while automatic pump samples provide the best coverage of storm events and adequate accuracy for fine sediment concentrations. Attempts to measure the gravel loads with a bag-type sampler are beginning.

At any given stage (depth), the observed concentrations vary depending on the passage of dunes, sediment waves, and watershed/channel conditions. A correction factor is introduced into the regression equations and updated at each average concentration observation to estimate sediment yields during observed flows. The uncorrected regression equations may be used to estimate expected yields by unmeasured or hypothetical flows in the channels.

INTRODUCTION

The Goodwin Creek Research Watershed is an 8.26 sq. mi. (21.4 sq. km.) instrumented watershed that is operated by the USDA Sedimentation Laboratory in Oxford, MS. This watershed includes fourteen gaging stations in the form of supercritical flumes. These structures serve dual functions as grade control and gaging structures but were designed primarily to facilitate measurements of flow rates and sediment transport rates. A detailed description of the watershed and its instrumentation is presented in a companion report [1]. This paper describes the types of sediment load measurements, analysis methods, and preliminary sediment ratings for the Goodwin Creek data.

Background:

Determining or predicting the quantities of sediment that may be delivered by flows in a stream channel is important for a number of reasons. First, the sediment delivered to a stream reach represents net erosion from the upstream areas - from either the watershed surface or the boundary of the channel itself. Second, the sediment conveyed downstream fills reservoirs, interferes with navigation, degrades water supplies, etc. And,

third, it represents material that must be handled by the local channel.

The concept of a "capacity" load of sediment must be viewed in some statistical average sense. Certainly, for the spectrum of flows and erosion rates that may occur in a watershed, the loads and flows cannot all be considered to be at capacity or in equilibrium. The existence of a bed of sand or gravel in a stream actually represents a source or sink of significant amounts of sediment to compensate for some deviation between supply and capacity. Bed forms and associated resistance changes also tend to compensate for variations in sediment concentration but may also contribute to variations in sediment load measurements [2,3]. Accelerating flow over the back of a dune entrains sediment causing an increase in the transport rate in the downstream direction. Deposition on the downstream face of the dune then reduces the transport rate to a minimum value at the dune trough to be followed again by a maximum transport rate at the next crest. A measurement of the bed material load associated with a dune crest will give a local load much higher than the "average" and the converse is true of measurements relative to the trough.

The silt and clay sediment is essentially uniformly distributed throughout the flow; therefore, it can be measured quite accurately by any reasonable sampling system. Since it is usually transported as "wash load", no model can be reliably used to estimate the fine load without calibration with actual data.

Numerous relationships have been proposed to estimate the transport rates of sand material from knowledge of the sediment properties and hydraulic factors of the flow. Some give reasonable estimates in certain applications but unreliable results for others. Because of the large differences between predictions, no predictor can be reliably applied without verification of its applicability by reliable measurements. Thus, obtaining reliable measurements of the sediment loads at each station constitutes a primary aspect of the Goodwin Creek data collection program.

SAMPLING EQUIPMENT

The supercritical flow structures of Goodwin Creek are "V" bottom concrete structures. The three largest structures have 5:1 side slopes for 15 ft. (4.6 m.) each side of center and 2:1 side slopes thereafter. The other structures have constant 2:1 sideslopes. Each of the structures has a 4 percent longitudinal slope which gives supercritical flow for all discharges that may be encountered and establishes fixed relationships between depth of flow and flow discharge rate within the supercritical

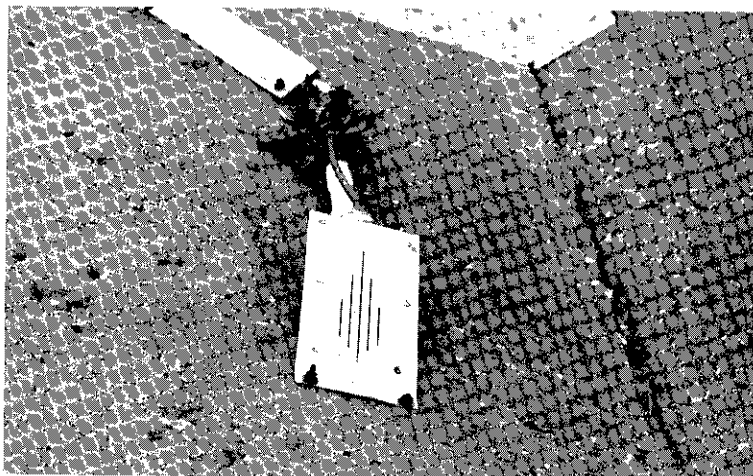
section. The high flow velocities also inhibit sediment deposition and provide a bottom reference for load measurements

Four methods of sediment load measurement are being used extensively within the structures. Figure 1 depicts the four sampler/intake configurations. The automatic samplers withdraw samples from a fixed point within the structure at set time intervals while the intake is covered by the flow. The intakes are mounted on the concrete surface either 0.5 ft. (0.15 m) or 1.0 ft. (0.30 m.) above the bottom of the "V". The intake consists of a sloping ramp to shed trash with 1/8-inch (.32 cm.) slots to exclude larger particles that might damage the pumps. An automatic sampler is installed at each of the fourteen channel stations as well as at two other single - source areas.

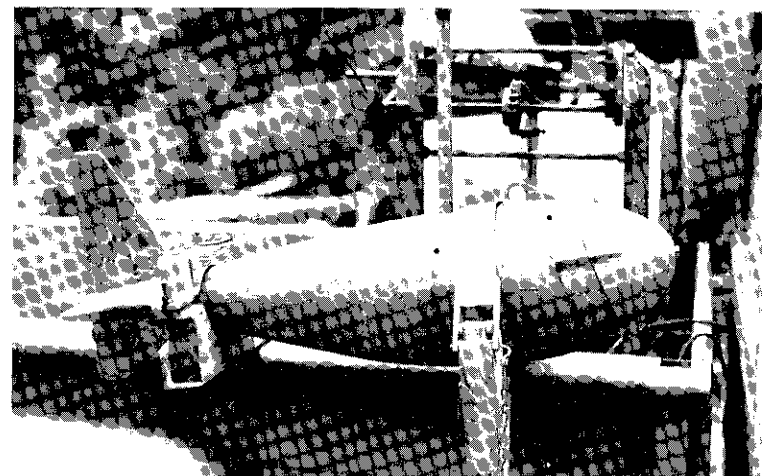
The automatic samplers include two different pumping systems. One of these uses the U.S. PS-69 automatic sampler with its standard pumps. The other uses a jet-assisted pneumatic pump along with a Chickasha sediment sampler [4]. The pneumatic pump uses dual diaphragms and check valves so that few moving parts are subjected to abrasive sediment. Adding a standard water well jet foot to the sampler intake and diverting part of the pump discharge through a pipe tee with a sediment diversion screen to supply the jet greatly reduces the suction head on the pump and improves pumping efficiency. It also allows the pump to operate with lower sand concentrations than the sample flow, since the jet supply is a circulating quantity of sand-free water. This pump system is better suited for high suction lifts than the PS-69, but the number of samples that can be collected without attention in the Chickasha device is only 28 versus 64 for the PS-69 system.

Since the automatic sampler intakes are located at fixed points for which relations of the sand concentration in the samples to the discharge - average sand concentration is unknown, manual depth integrated samples (equal-transit-rate, ETR, samples at selected verticals or single-vertical samples) are serving as the primary means of determining sand loads. The first of these manual measurement techniques utilized a cable- supported U.S. P-63 sampler to measure the average concentration of suspended sands. The cable mounted system is being replaced by a boom-mounted carriage that traverses a U.S. DH-48 sampler through the flow nape at the end of the structures. Since this boom-mounted system samples the entire depth of flow, samples collected with this system are referred to as "total load" samples. The cable- mounted system could not sample closer than about 0.3 ft. (0.1 m) of the structure surface and lacked the needed horizontal control.

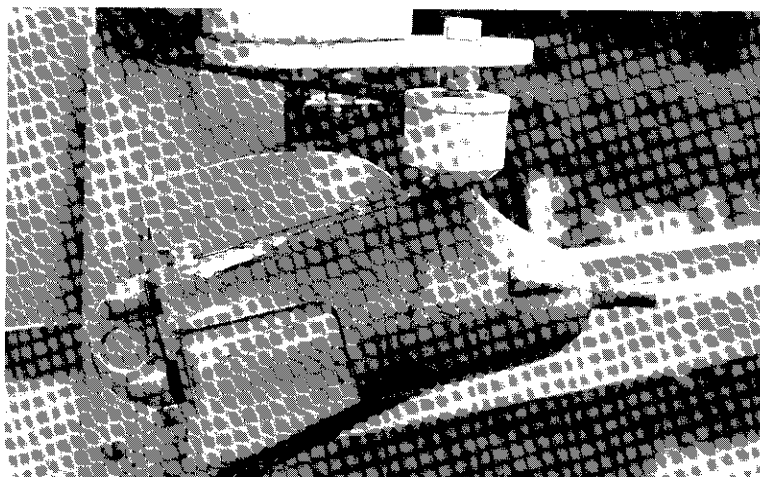
The "total load" samples are actually samples of the total loads of sands and finer sediments since the DH-48 intake is a



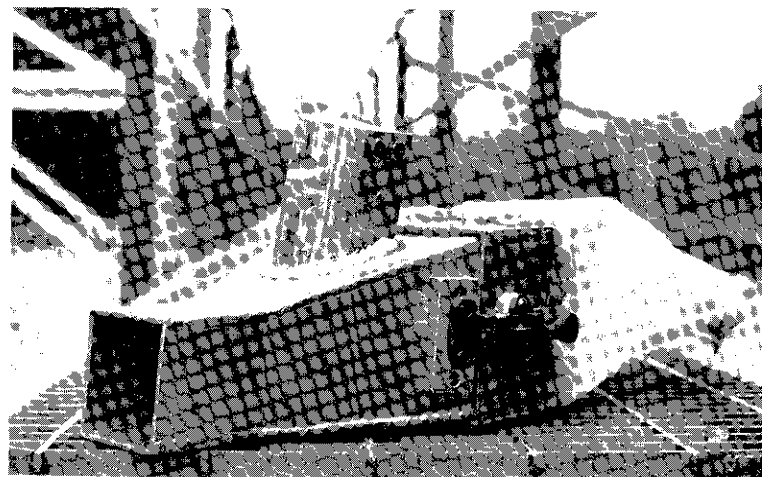
Automatic Sampler Intake



US P-63 Sampler



US DH-48 Sampler



Modified Helley-Smith Sampler

Fig. 1: Sampler intakes used on Goodwin Creek, MS

1/4-inch (.64 cm.) nozzle. A bag-type Helley-Smith sampler [5] has been adapted to a similar boom support at the upstream entrance to the critical flow section to collect samples of gravel in transport. The standard 3x3-inch (7.6x7.6 cm) intake was deformed to a parallelogram to fit the surface of the structures. This sampler rests on the bottom for timed periods at distributed points across the structure to estimate the gravel load and the distribution of gravel in transport across the structure.

ANALYSIS TECHNIQUES

Individual sediment samples are analyzed for concentration of fine sediment, sands, and gravel as appropriate. The total load samples and gravel samples are analyzed for particle size distribution. Each sample is cataloged on the laboratory computer according to time of collection and gage height (flow depth) in the structure. Average concentration of sand or fine sediment is determined from a composite of ETR samples or by weighting single-vertical samples according to their respective flow areas. Gravel samples will be treated as single-vertical samples when enough are available for interpretation.

The flow discharge rate through the structures is determined from the theoretical ratings with gage height which agrees well with flume model tests and with actual prototype measurements. The product of the average sediment concentration C and discharge rate Q then provides estimates of the sediment transport rates.

Since the average energy slope of a long channel reach remains essentially constant, the average hydraulic variables of the flow such as shear stress, velocity, and Froude number should all be related to the flow rate and consequently to the depth H in the structures. Neglecting deviations from equilibrium, the bed material load should depend on the hydraulic variables and consequently on H . A number of different functional forms were tried in regression analyses to relate observed C and H values. Over-representation of certain depth ranges tended to bias the results, so averages of concentrations within each one-fourth foot interval of depth were computed and used to establish the regressions. Of the functional forms that were tried, a power function of the form of Eq. 1 consistently gave the best fit in terms of the squared correlation coefficient RSQ .

$$C_i = D_i H^{E_1} \quad (1)$$

The subscript i denotes a particular sediment size class (sands or fines.) Better correlations were obtained by

considering separate regression equations for rising and falling stages for the fine sediment. Insufficient rising- side hand samples are available to warrant this distinction for sands.

Appreciable scatter exists for the concentration values of each depth interval. The standard deviations of concentrations observed within each 1/4-ft. and the standard error of estimate are considered to quantify the data scatter about the mean concentration of a depth interval or about the assumed regression equation, respectively. Hence, the squared correlation coefficient, which is related to the ratio of the total standard deviation to the total standard error of estimate is considered to indicate how well the regression equation fits the 1/4-ft. interval averages. Regression equations such as Eq. 1 should be considered to represent the long-term average concentration. Scatter about regression equations are usually viewed as errors but this is not true for concentration data. Concentrations are observed to deviate appreciably and consistently from the average curve for entire storm periods or even for several storm events, but no consistency to the deviations have been identified. Any attempt to use the regression equation to estimate the actual sediment delivery for specific time periods or events must include a correction for the current deviation as reflected by recent samples. This correction is facilitated by adding a shift coefficient K to Eq. 1 to account for the deviation.

$$C_i = K D_i H^{E_i} \quad (2)$$

For an observed concentration/depth pair (C_i, H) , and regression values of D_i and E_i , the value of K can be determined and used to revise concentration estimates between samples. For changing K values, a timewise interpolation of the K values for successive samples is used.

A few erroneous concentration values are present in any data set. Temporary clogging of sampling lines or nozzles can give unreasonably low values followed by unreasonably high concentrations as the line is purged. A filtering technique was defined to delete obviously erroneous samples. If an observed concentration gives a K value greater than 3. or less than 1/3 (deviating by a factor of 3 from the regression eqn.) and if it deviates by more than a factor of 3 from the average of adjacent concentration observations, it is deleted from the analysis. Although only a few but extreme values are deleted, a revised regression equation is then computed and used for load estimates.

RESULTS

More data have been obtained from the automatic samples so these have served as the primary basis for the results obtained thus far. Table 1 summarizes the regression computations based on Eq. 1 for the fine sediment data of automatic and manual samples.

Table 1: Regression Parameters - Fine Sediments -
Goodwin Creek, MS

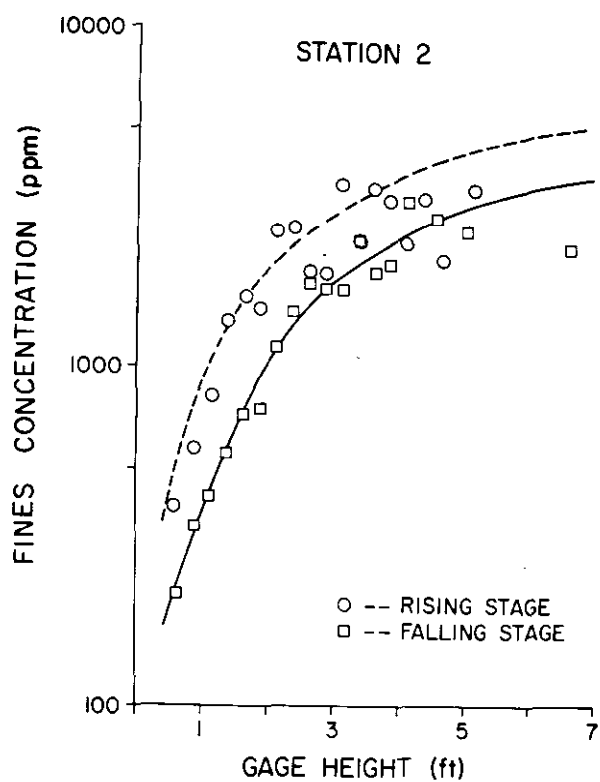
Station	Rising Hydrograph			Falling Hydrograph		
	D	E	RSQ	D	E	RSQ
1	1613	0.636	0.649	631	0.991	0.873
2	805	0.946	0.882	419	1.20	0.929
3	792	0.464	0.526	454	0.713	0.702
4	868	1.026	0.841	632	1.16	0.863
5	1156	0.488	0.643	601	0.870	0.897
6	1190	0.826	0.836	758	1.12	0.910
7	2789	0.563	0.770	870	1.12	0.836
8	1018	0.877	0.624	600	0.999	0.779
9	6504	0.281	0.085	2937	0.689	0.409
10	56	-0.552	0.424	63	0.742	-----
11	1649	-0.043	0.004	600	1.11	0.931
12	1442	1.10	0.811	599	1.64	0.986
13	2963	0.577	0.560	1030	0.631	0.357
14	794	1.31	0.667	518	1.35	0.896

Figure 2A illustrates typical regression curves for Station 2 along with the average fine sediment concentrations for 1/4-ft. depth intervals. The average concentration values deviate from the regression equation in certain depth ranges, but in view of the large standard deviations, this deviation is considered not to be significant. The standard deviations of concentrations for each 1/4-ft. depth interval were on the order of the average concentrations themselves. Higher concentrations for the rising portions of hydrographs is typical of most observations.

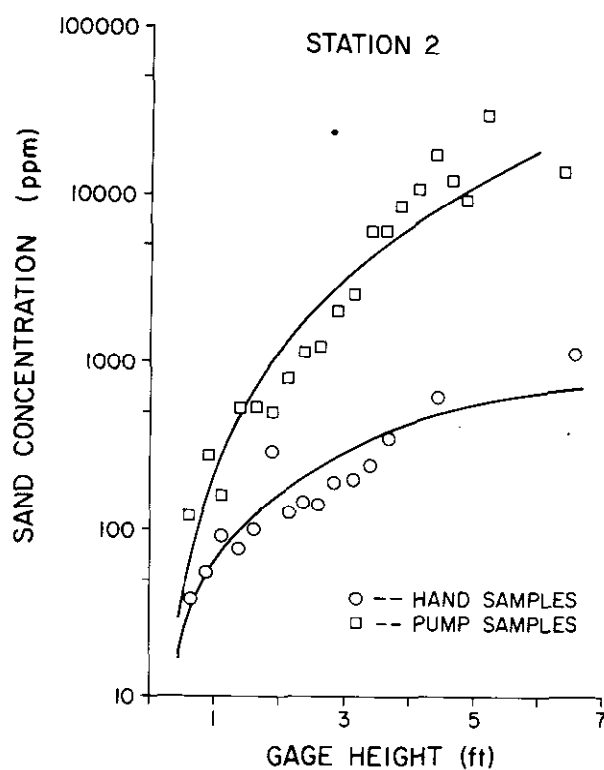
Table 2 summarizes the regression information for the sand fractions of the automatic samples and hand samples. Although the correlation for some of the stations is at least fair, these single-point samples do not correspond directly to the average concentrations of sand over the entire crosssection. For significant depths, the intakes for the automatic samplers are near the bottom where sand concentrations are high. Thus, a sampling "efficiency" relationship must be defined to convert these concentrations to true average concentrations. Since each sampler intake's relative elevation in the flow depends on the

Table 2: Regression Parameters - Sands - Goodwin Creek, MS

Station	Automatic Samples			Hand Samples		
	D	E	RSQ	D	E	RSQ
1	260	2.97	0.775	40	1.72	0.928
2	211	2.53	0.911	62	1.32	0.862
3	466	1.82	0.902	148	1.04	0.553
4	176	2.94	0.955			
5	340	1.74	0.683	161	0.766	0.508
6	632	2.44	0.978			
7	157	3.02	0.983			
8	44	2.94	0.885			
9	5860	1.11	0.613			
10	11	-1.35	0.579			
11	315	2.87	0.973			
12	80	2.04	0.921			
13	770	1.28	0.293	339	0.915	0.557
14	872	1.40	0.720			



2A



2B

Fig. 2: Average sediment rating curves for Goodwin Creek, MS

depth H, the sampling efficiency may also depend on H.

The manual sampling techniques are considered to give the best estimate of the average concentration of sand-size sediment. Figure 2B depicts the depth-interval average concentrations of sand in the automatic samples and in the cable-suspended and boom-mounted samples (hand samples) for Station 2. For this station the ratio of the regression equations varies from about 3.4 for a depth of 1.0 ft. to a ratio of 34. for a depth of 6 ft. If these ratios prove to vary consistently with H, then sand load estimates may be obtained from the automatic samples; however, these load estimates and the regression factors of Table 2 should be considered very preliminary.

Attempts to measure the gravel fractions of the sediment load in Goodwin Creek are just beginning. Some preliminary data using a temporary box trap at the end of Structure 13 gave widely varying results concerning relative transport rates of sand and gravel. During periods of high sand load, little gravel was transported but for similar flows with a smaller sand load, appreciable gravel was collected in the trap. The small watershed above Station 13 gives quick response to cover and tillage operations therein. High sand delivery rates seem to cover the gravel in the channel bed with sand effectively shielding it from the flow. As the sand is removed during periods of low sand delivery, gravel particles are again exposed as part of the bed material and their transport resumes.

CONCLUDING REMARKS

Procedures are being developed to obtain "best" estimates of the sediment transport rates in the Goodwin Creek Research Watershed. This has involved developing sampling equipment compatible with the supercritical flow structures and developing analysis procedures to extract the most meaningful information from the data. A manual sampling system provides integrated samples and the most accurate sand concentration measurements. The automatic pump samplers provide better coverage of storm events, adequate accuracy in concentration of fine sediments, and sand concentrations that may be related to the average concentration of the total flow. Adaptations of Helley-Smith bag-type samplers are being used in an attempt to measure the transport rates of gravel.

During periods of adequate sample coverage, estimates of the actual sediment load may be obtained by successive updates of a shift factor to account for consistent deviations of observed concentration from the average curve. Concentrations for time periods between samples may be estimated from the shifted regression. The average curve without the shift factor can be

used to obtain the expected yield of sediment by observed or hypothetical flow events. It may also be convoluted with flow duration relationships to give projections of long-term yields.

The sediment load measurement program in Goodwin Creek is ongoing with an expected long duration. The large deviations in observed concentration coupled with their long durations require extensive data over long time periods to define average relationships and statistical parameters associated with the transport processes. Seasonal changes, land use changes, and extreme events can all introduce perturbations - perturbations that we would like to quantify. The long range quest of the program is to develop relationships that will not only predict long-term average transport processes but also the effects of changes in watershed conditions on the transport quantities and associated channel stability. Existing transport relationships may be applicable or adaptable to describe the transport processes, but these relationships have all been developed for continuous size distributions of sediment. The varying, bimodal distributions of sand and gravel in the bed material in Goodwin Creek represents a need for transport research in both the field and the laboratory.

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A LABORATORY STUDY OF SEDIMENT TRANSPORT UNDER ICE

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ABSTRACT

The influence of an ice cover on flow in an alluvial channel involves a complex interaction between the ice cover, fluid flow, fluid properties, sediment and bedforms. The presence of an ice cover changes an open channel into a form of closed conduit, which in turn affects the magnitude and distribution of velocities as well as the boundary and internal shear stresses. In addition, the colder water temperatures affect fluid properties important to the transport of sediment. Further, the boundary roughness is not constant since the interaction between the ice cover and bed may induce a change in boundary form with time and location.

This paper will discuss the results obtained in a 36-m-long refrigerated, recirculating flume that is 1.2 m wide and 0.6 m deep. The bed was composed of a uniform 0.45-mm quartz sand. For a given discharge the flume was first operated as an open channel until essentially uniform equilibrium conditions were attained. Runs were then repeated with the addition of simulated smooth and rough ice covers at the same discharge and flume slope, allowing depth to vary. The test results were used to examine the effects of an ice cover on bedform, resistance to flow and total sediment transport.

INTRODUCTION

The formation of ice on rivers and lakes is an annual occurrence in over half the United States. The duration and extent of ice formation vary greatly, but for a significant portion of the country, ice and its effects are important considerations in understanding fluvial systems and in the design, operation and maintenance of associated hydraulic structures. The influence of an ice cover on an alluvial system involves a complex interaction between the ice cover, fluid flow, fluid properties, sediment and bed forms. The presence of an ice cover roughly doubles the wetted perimeter of a wide channel, which in turn affects the magnitude and distribution of velocities and the boundary and internal shear stresses. In addition, the lower water temperatures associated with the presence of ice alter fluid properties, such as viscosity, which can affect resistance to flow and transport capacity. As an added element of complexity, the boundary roughness is not constant since the roughness of both the sediment bed and ice cover may vary with composition, form, and time.

LITERATURE REVIEW

The subject of sediment discharge in ice-covered streams has not been widely addressed to date. The techniques of sediment discharge measurements for periods of ice cover are quite different from those made during open water conditions. As discussed by Tywoniuk and Fowler (1972), there are problems not only with the functioning of manpower and equipment during severe winter conditions, but also with the relatively unknown nature of the interaction of ice with sediment transport and the variability of ice conditions that affect this interaction. While some studies have evaluated the resistance to flow for ice-

covered streams, there has been little documentation of sediment transport or bed forms. Due to the uncertain accuracy of winter measurements, even the limited field data available are subject to question.

Sayre and Song (1979) have investigated the effects of ice covers on sediment transport and bed forms in the laboratory. Because of the reduction in flow velocity and bed shear stress due to the added resistance of the ice cover, they concluded that the presence of an ice cover will tend to reduce sediment discharge. Experimentally, they found that total sediment discharge decreases sharply relative to open water conditions, and both wave height and wave length of ripples are decreased. As expected the added boundary area decreased velocities and increased depth. In their work, Sayre and Song used floating hinged plywood to simulate the presence of a smooth ice cover, with Masonite strips added to simulate a rough ice cover.

Akyeampong (1971) experimentally studied the local sediment transport rate in periodically non-uniform flows in a wavy boundary conduit. His experimental setup consisted of a recirculating pipe loop with a rectangular working section. The top boundary of the working section had a sinusoidal longitudinal profile. The objective was to study the phase shift between local sediment transport rate and the velocity perturbation due to the wavy top boundary. Akyeampong documented a phase shift between both local velocity and sediment transport and between the bed form and upper boundary. Relating this to the effects of an irregular ice cover, Akyeampong found that surface boundary waves have a strong effect on velocity profiles and distributions of pressure and shear stress. Also changes in water temperature have significant effects on the phase shifts, and the concentration of sediment at any elevation above the bed varies along the channel in a sinuous manner.

More recently, the effect of ice on suspended sediment transport has been studied analytically by Krishnappan (1983) using a $k-\epsilon$ turbulence model. He found that suspended sediment concentration should decrease as the ratio of ice cover roughness to bed roughness increases. Subsequently, Lau and Krishnappan (1985) presented the results of a laboratory flume study using glass beads for sediment and a plywood and plastic composite ice cover. Again, they used a $k-\epsilon$ turbulence model to evaluate flow parameters and assumed that the flow below a hypothetical zero shear plane (corresponding to the point of maximum velocity in any vertical) could be treated as a free surface channel. They concluded that the addition of an ice cover will drastically reduce sediment discharge, but bed load and bed form geometry can be predicted using free surface relations if the lower layer hypothesis is used. Although both bed load and suspended load were reduced with the ice cover, the effect was much more pronounced for suspended load.

ANALYTICAL CONSIDERATIONS

We can make a simplified analysis of the effects of adding a surface boundary to a uniform flow system using the commonly employed Manning relation (metric units):

$$Q = \frac{1}{n} A R^{2/3} S^{1/2} \quad (1)$$

where Q is the water discharge, n is Manning's roughness coefficient, A is the cross-sectional area, R is the hydraulic radius and S is the energy slope.

Barring long-term changes in channel slope, the addition of a uniform floating ice cover should not affect the energy slope. The hydraulic radius, R, is dramatically affected, however. If we assume a wide, open channel we may simplify R as

$$R = \frac{W \cdot D}{W + 2D} \approx \frac{WD}{W} = D \quad (2)$$

where W is the channel width and D is the depth. In the case of an ice-covered channel, however, the same simplification results in

$$R = \frac{W D}{2W + 2D} \approx \frac{D}{2} \quad (3)$$

and the hydraulic radius is essentially halved with the addition of an ice cover. If we now consider the addition of an ice cover to a channel passing a constant discharge we can show that $Q_o = Q_I$ so that

$$\frac{A_o D_o^{2/3}}{n_o} = \frac{A_I (D_I/2)^{2/3}}{n_I} \quad (4)$$

where the subscript o indicates the open channel case, I indicates the presence of an ice cover and n_I is a composite Manning's n. If we assume a steep-sided channel so that width does not vary appreciably with depth,

$$\frac{D_o^{5/3}}{n_o} = 0.63 \frac{D_I^{5/3}}{n_I} \quad (5)$$

$$\text{and } D_I = 1.32 D_o \left[\frac{n_I}{n_o} \right]^{0.6} \quad (6)$$

For the case of a rigid bed such that bed roughness does not change in response to ice cover effects, Carey's (1966) calculations indicate $0.73 < n_I/n_o < 1.37$ such that:

$$1.09 < \frac{D_I}{D_o} < 1.59. \quad (7)$$

Through a similar analysis it can be shown that the ratio of velocities is

$$0.627 < \frac{V_I}{V_o} < 0.915 \quad (8)$$

and assuming that the boundary shear, τ_b , can be represented as $\tau_b = \gamma R S$ (where γ is the specific weight of water):

$$0.55 < \frac{\tau_{bI}}{\tau_{bo}} < 0.797. \quad (9)$$

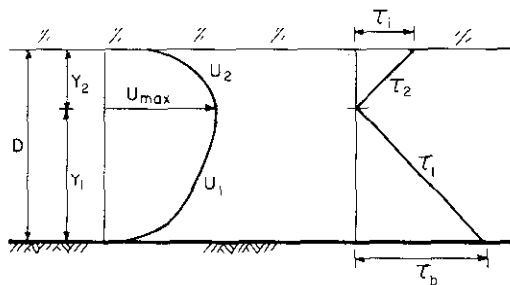


Figure 1. Velocity and shear stress distributions for an ice-covered channel.

Thus, this analysis shows that a uniform floating ice cover could substantially increase flow depth while decreasing flow velocity and average boundary shear. In an alluvial system the changes are considerably more complex because of the ability for the bed to adjust its form and resistance in response to the hydraulic changes imposed by the ice cover.

When considering the internal characteristics of an ice-covered flow, it has been commonly assumed that the flow may be split along a hypothetical plane of zero shear passing through the point of maximum velocity in any vertical profile. This results in an upper and lower layer, as shown in Figure 1, which can be treated separately. More recently Tatinclaux and Gogus (1983) have shown that this hypothetical position of zero shear is displaced somewhat toward the rougher boundary because of the influence of Reynolds stresses. Unless the roughnesses of the two surfaces are quite different, however, this effect is small.

Using this two-layered approach (at least when bed load predominates), we can envision that the lower layer is in effect a new and independent "free surface." The adequacy of this assumption will be examined by comparison with the preliminary data sets obtained in a laboratory flume.

EXPERIMENTAL INVESTIGATION

Tests were conducted in a 36-m-long refrigerated, recirculating flume that is 1.2 m wide and 0.6 m deep. The bed was a uniform quartz sand with a median diameter $d_{50} = 0.45$ mm. Five sets of runs at four water discharges were completed, all at room temperature, approximately 19°C.

For a given discharge the flume was allowed to operate until an equilibrium uniform (at least in the statistical sense) flow was established having a mean depth of 15 cm. After all measurements were completed, the same discharge and bed slope were retained but a simulated ice cover made of extruded styrofoam was added. Subsequently, a run was conducted with the "ice cover" roughened by covering the underside with a plastic "bubble" packing material. Manning's n for the smooth and rough covers were on the order of 0.016 and 0.026 respectively.

Measurements of bed slope elevation and geometry were made using an ultrasonic transducer mounted on a motorized carriage that could traverse the length of the flume. Water surface elevations and slopes were measured using a series of piezometer taps along the flume bed and read by point gage.

Total sediment discharge was measured using a 1-cm-wide full-depth slot sampler mounted to intercept the free overfall in the flume tailbox. This sampler was

mounted on a carriage system so that it could traverse the full width of the flume. Suspended sediment samples were collected with a 0.95-cm-diameter nozzle with withdrawal rate regulated by siphon adjustment.

Velocity measurements were obtained with a two-directional electromagnetic velocity meter with a 1.25-cm-diameter spherical transducer. Total discharge was monitored by an in-line electromagnetic flow meter with a feed-back system allowing continuous control.

RESULTS

The major variables examined in this paper include bed form height, wavelength, Manning's roughness, and sediment discharge. Since all runs were conducted with the same uniform sand at the same temperature and with the same open water depth, the results are directly comparable and are presented in simple dimensional form. Further, for any given discharge, the flume slope remained unchanged for the open water, smooth ice and rough ice cases. The slope of the bed did change, however, becoming steeper with the addition of the ice cover.

In the experiments, bed forms were limited to the ripple and dune regimes. Although currently available bed form predictors could at least be conceptually applied to an ice-covered flow, it is not yet clear if the full flow depth or lower layer thickness should be used in predictors using the Froude number Fr , relative roughness D/d , or stream power τU .

In the following discussion both approaches are used. The full depth approach is perhaps more indicative of what happens in terms of the channel as a whole, while the lower layer approach treats the near bed region as an independent flow system.

Bed Form Height

Figure 2 presents measurements of average bed form height relative to stream power based on the full flow depth ($\tau_0 U = \gamma D S U$). In this form the data indicate that for the same value of stream power, bed form heights would decrease with the addition of an ice cover. Subsequent roughening of the cover, however, causes a slight increase. If the energy slope of the system did not change with the addition of an ice cover, then total stream power would remain constant for a given discharge since a change in depth would be compensated for by an opposing change in velocity. Actually the sand bed slope does change, becoming steeper with the addition of an ice cover or roughening of the cover. However, the same trends in bed form height are present in terms of constant stream power or discharge.

If Figure 2 had been plotted in terms of stream power based on the lower layer, it would have shown a substantial increase in bed form height with the addition of an ice cover. For a given point on the figure, all terms in the stream power relation would be constant with the exception of depth. For the smooth ice cover, depth (and thus stream power) would be decreased about 48%, and for the rough ice cover about 53%. No method of parameterization was found to collapse the open water and ice cover data onto a single curve, or to explain the observed data trends.

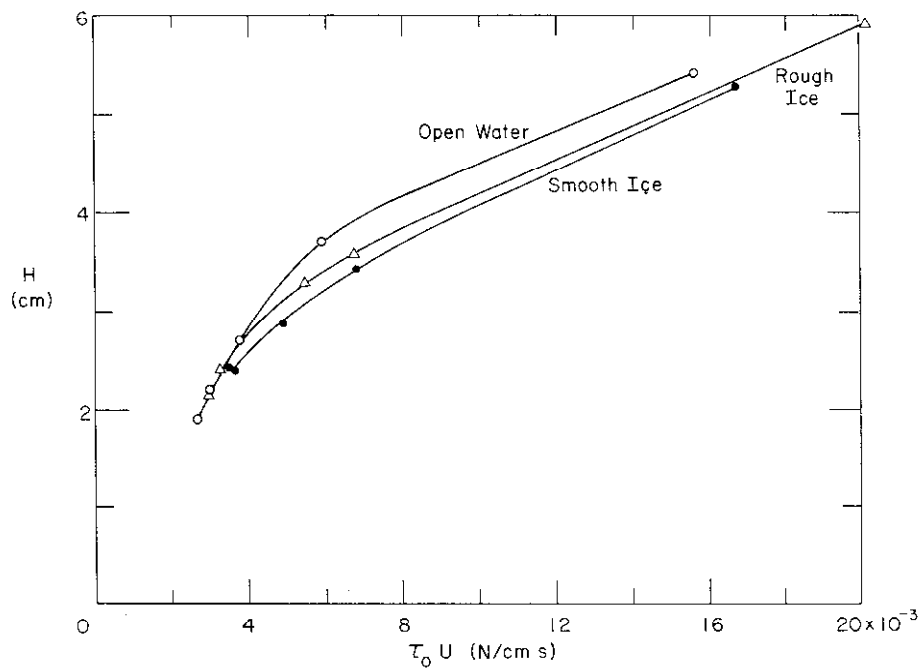


Figure 2. Bed form height vs stream power.

Bed Form Wavelength

As in the case of bed form height, the measured bed form wavelengths are compared in terms of total stream power. As shown in Figure 3 the addition of an ice cover (or its subsequent roughening) causes an increase in bed form length. If stream power were calculated in terms of the lower layer, even greater increases would be indicated.

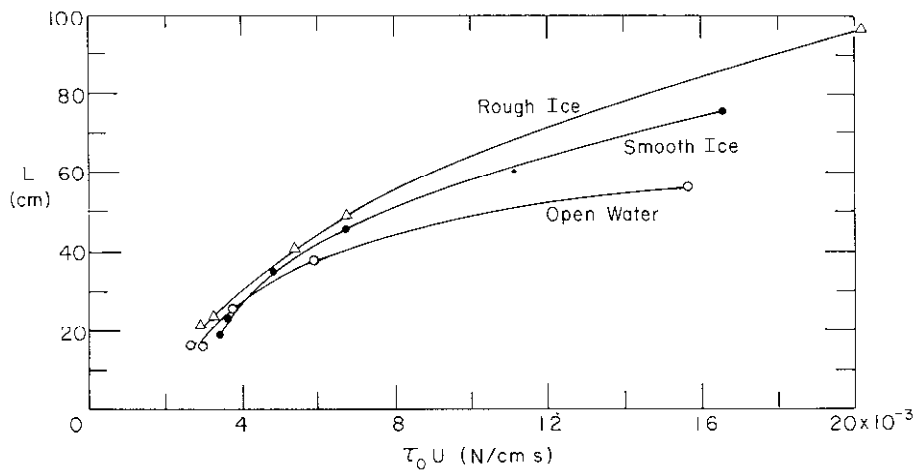


Figure 3. Bed form length vs stream power.

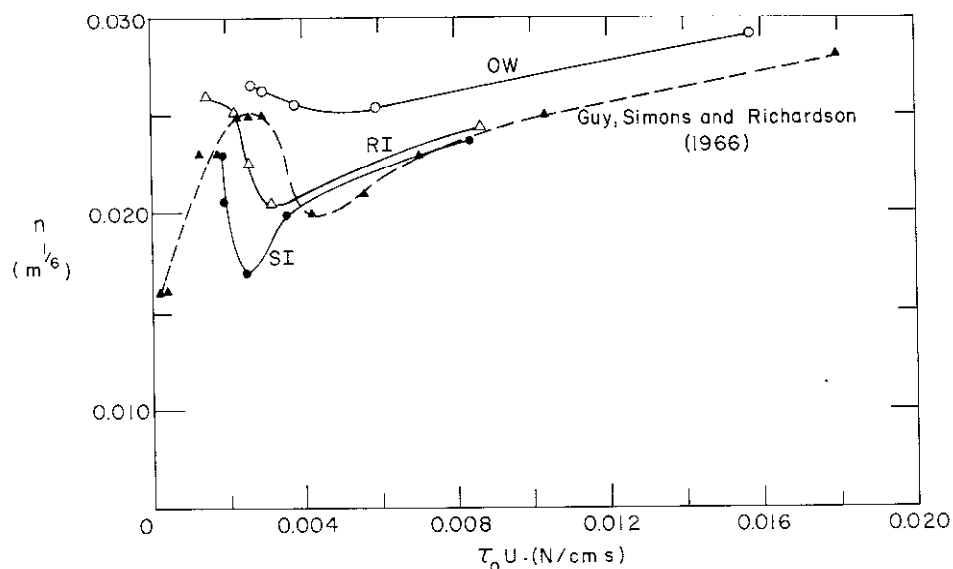


Figure 4. Manning's n vs stream power.

Resistance to Flow

Calculated Manning's n values for the sand bed are plotted in Figure 4 in terms of total stream power. For comparison, data for similar sand size and water depths reported by Guy et al. (1966) are also shown which cover a wider range of stream power. The minimum point at about $\tau_0 U = 0.0042$ in their data corresponds to a transition between ripples and dunes. The shift in this minimum point between the ice curves may indicate an ice effect on this transition.

It is interesting to note that, for the same value of stream power, roughness is significantly reduced with the addition of an ice cover, but an increase in the ice cover roughness results in an increase in the bed roughness. This is analogous to the results for bed form height shown in Figure 2 where the smooth ice cover results in the lowest wave heights. Using stream power based on the lower layer would shift the ice cover curves toward smaller values, but the trends in roughness would be unaffected.

Sediment Discharge

Figure 5 shows the concentration of total sediment discharge with stream power. Although suspended sediment samples were collected, concentrations were small relative to bed load. Thus the quantities shown in Figure 5 were moved almost exclusively as bed load.

As in the case of bed form geometry we were unable to find a good parameter to correlate the transport data. In the case of Figure 5 where stream power is based on the total flow, the addition of a surface cover drastically reduces sediment transport for the same value of stream power. Roughening the ice cover does not significantly affect the relation. If we base stream power on the lower layer instead, we obtain substantially higher transport than for the open water case, but since roughening the ice cover increases transport it is apparent that we again lack a good parameter for correlating the data. The

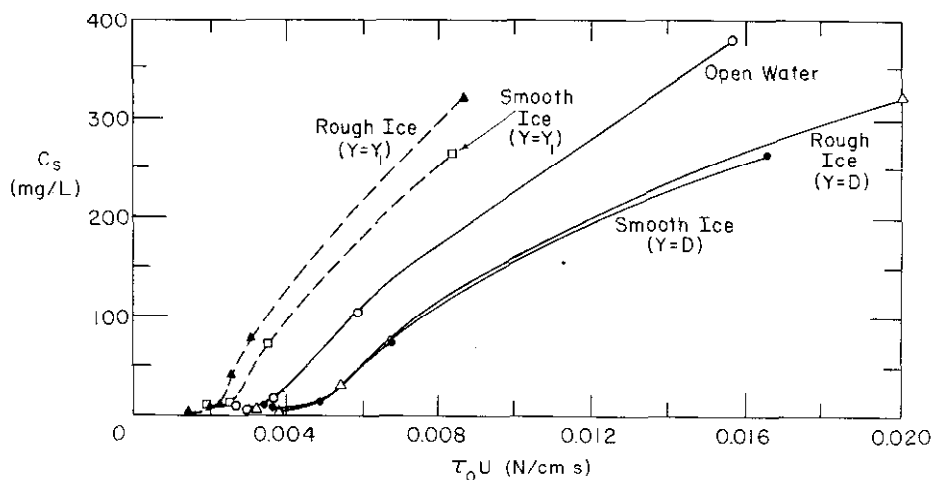


Figure 5. Sediment discharge concentration vs stream power.

indicated increase in sediment transport may be due to the way the data were treated. In terms of the total water discharge, sediment concentration is reduced by the addition of an ice cover.

SUMMARY AND CONCLUSIONS

A preliminary set of laboratory flume data was analyzed to examine the effect of an ice cover on sediment transport and resistance to flow. A simple analysis based on the Manning equation indicates that an ice cover would increase flow depth while decreasing mean velocity, boundary shear and sediment transport. A satisfactory parameter to correlate open water and ice cover tests was not found, and the data did not support the hypothesis that the flow below a defined plane of zero shear could be treated as an independent system.

Based on the characteristics of the total flow, the addition of an ice cover will increase water depth and reduce velocities and bed shear. The data indicate a reduction in sediment transport, bed roughness and bed form height, but an increase in bed form wavelength. Since further roughening the ice cover increases the bed form height and increases Manning's n , it appears that the parameters used in the correlation of open water to ice conditions were inadequate. The problem appears to center on an appropriate depth scale.

If we instead use the hypothesis of flow below a plane of zero shear (passing through points of maximum velocity in vertical profiles), we again find that correlations based on stream power are inadequate. In this case the addition of an ice cover would increase sediment concentration, bed form height and wavelength, but bed roughness would still be somewhat reduced.

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ACCURACY OF SEDIMENT DISCHARGE ESTIMATES

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ABSTRACT

The accuracy of various methods for estimating long-term suspended-sediment discharge was evaluated by comparing the discharge measured at a long-term daily sediment station with the discharge estimated by 24 different methods. The flow-duration, suspended-sediment transport curve procedure was the basis for all the methods. Differences between the methods were the types of curves, the dependent variable, and how the curves were determined. Types of curves included a single curve, two seasonal curves (dormant and growing seasons), and three size fraction curves (sand, silt, and clay). Relations between sediment concentration, sediment discharge, and water discharge were developed for each type of curve. The curves were computed by linear and nonlinear regressions or hand drawn through average sediment-concentration or sediment-discharge values within specified water discharge intervals. Daily suspended-sediment discharge records and particle-size determinations for the Schuylkill River at Manayunk, Pennsylvania, were used in the analyses. After comparing the discharges estimated by the 24 methods with the measured discharge, three of the more suitable methods were tested with data from other daily sediment stations. Five stations with drainage areas ranging from 10.2 to 24,100 square miles were used to evaluate the suitability of the methods for both smaller and larger streams. The linear regression and group average methods of relating sediment concentration or sediment discharge to water discharge produced suitable estimates of long-term discharge. None of the methods produced good estimates for the small streams.

INTRODUCTION

The measurement of suspended sediment transported by streams and rivers is an important component of any assessment of river quality. Suspended-sediment data provide information for evaluating water-treatment costs, flood-water and sediment-storage requirements for reservoirs, dredging requirements for waterways and harbors, and the quality of aquatic resources. These data also are important for assessing the distribution and transport of toxic substances because many trace metals, pesticides, and other organic compounds are sorbed to particulate material in water.

Much of the available suspended-sediment data has been collected at daily sediment stations; this involves the collection and analyses of several hundred samples each year. Daily stations provide a wealth of information; however, they are very expensive to maintain and operate. Costs include manpower and equipment for collecting samples, transportation of samples to a laboratory, laboratory analyses, and interpretation and computation of sediment concentrations and discharges.

Several different methods based on short-term or periodic sampling have been developed to estimate the long-term suspended-sediment discharges of streams. One of the most popular methods is the flow-duration, sediment-transport curve procedure described by Miller (1951) and Colby (1956). This method involves

defining a relation between water discharge and suspended-sediment concentration or discharge (transport curves) and then computing the long-term sediment discharge from the long-term water discharge records. Various investigators have used different types of sediment transport curves such as monthly, seasonal, or annual curves. Transport curves also have been developed for different particle sizes of suspended sediment.

Purpose and Scope

The purpose of this paper is to compare different procedures for estimating long-term suspended-sediment discharge using data from one moderate-sized river and to evaluate the more suitable methods using data from both small streams and large rivers. The daily suspended-sediment record for the Schuylkill River at Manayunk, Pa., was used to compare the different methods. Three of the more suitable methods were then tested with the records of five other streams in Pennsylvania and Delaware (fig. 1 and table 1). The drainage basins of the streams range from 10.2 to 24,100 mi² (square miles).

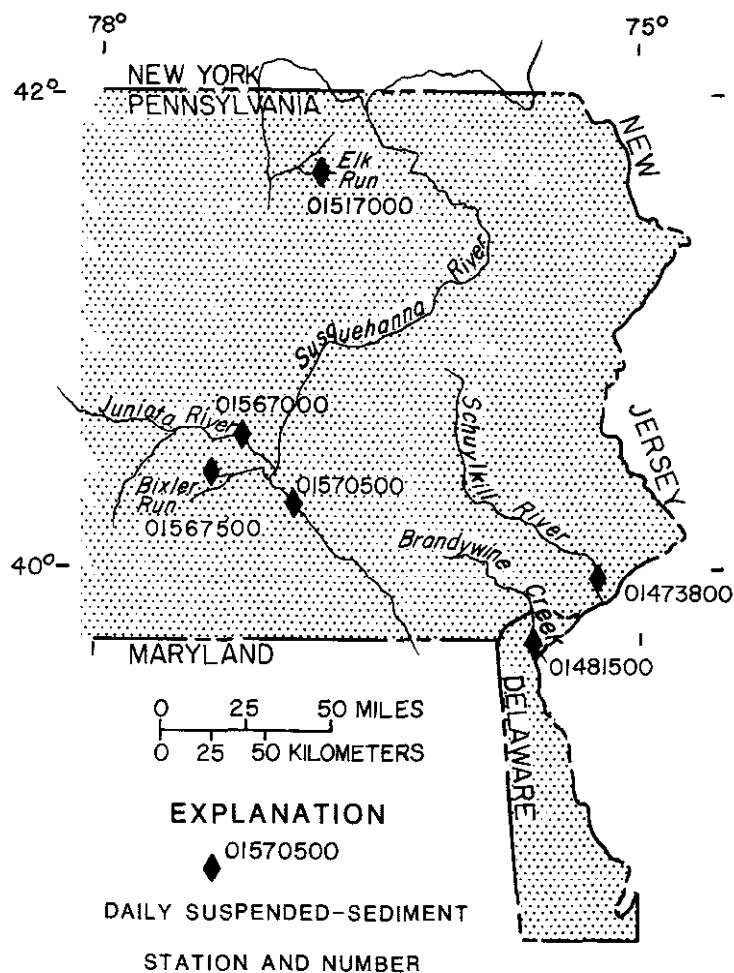


Fig. 1. Location of daily suspended-sediment discharge stations used in the study.

Table 1. Predominant land use, rock types, and mean water discharge of the study basins

Station number	Station name	Drainage area (mi ²)	Predominant land use	Principal rock types	Average annual water discharge [(ft ³ /s)/mi ²]
01473800	Schuylkill River at Manayunk, Pa.	1,810	Agriculture-forest	Shale, sandstone, siltstone, and quartzite	1.55
01481500	Brandywine Creek at Wilmington, Del.	314	Agriculture-urban	Granite, gneiss, and shist	1.55
01517000	Elk Run near Mainesburg, Pa.	10.2	Agriculture	Sandstone, siltstone, and shale	1.08
01567000	Juniata River at Newport, Pa.	3,354	Forest	Interbedded shale, sandstone, and limestone	1.28
01567500	Bixler Run near Loysville, Pa.	15.0	Agriculture-forest	Limestone mixed with shale and sandstone	1.24
01570500	Susquehanna River at Harrisburg, Pa.	24,100	Forest-agriculture	Interbedded sandstone, shale, and limestone	1.43

Description of Study Basins

The land use, geology, and mean streamflow of the study basins are summarized in table 1. The Schuylkill River at Manayunk drains 1,810 mi², and land uses range from anthracite mining in the forested mountains in the headwaters to extensive agriculture in the middle basin and urban areas in the lower basin. More than 9.9 million tons of suspended sediment were discharged by the river during the base period for this study, October 1, 1953, to September 30, 1979. Annual suspended-sediment discharges ranged from 107,000 tons in 1966 to 996,000 tons in 1972; the yearly mean water discharge ranged from 1,010 to 4,520 ft³/s (cubic feet per second) during the same period. The suspended sediment transported during the 26-year period was 6 percent sand, 54 percent silt, and 40 percent clay.

The other drainage basins differ considerably from the Schuylkill River basin in either their size, land use, geology, or streamflow characteristics. Brandywine Creek is about one-fifth the size of the Schuylkill River at Manayunk; Elk Run and Bixler Run are two orders of magnitude smaller; and the Juniata and Susquehanna Rivers are 2 and 13 times larger, respectively. Predominant land uses in the Schuylkill River basin are agriculture and forest; the other basins are similar except that Brandywine Creek has a higher percentage of urban land and the

Juniata and Susquehanna River basins are more forested. The Schuylkill River basin is underlain by shales, sandstones, siltstones, and quartzite; the other basins have similar geology except the Brandywine Creek basin, which is underlain by igneous and metamorphic rocks, and Bixler Run, which is underlain by limestone mixed with shale and sandstone. Water yields for the basins range from 1.08 (ft³/s)/mi² (cubic feet per second per square mile) for Elk Run to 1.55 (ft³/s)/mi² for the Schuylkill River and Brandywine Creek.

METHODS FOR ESTIMATING LONG-TERM SEDIMENT DISCHARGE

Twenty-four different methods for computing long-term suspended-sediment discharges were evaluated by comparing the discharges estimated by each method with the measured suspended-sediment discharge from October 1, 1953, to September 30, 1979. Each method used a flow-duration curve representative of long-term stream-flow characteristics and a relation between water discharge and suspended-sediment concentration (or discharge) to compute the long-term suspended-sediment discharge (Miller, 1951, and Colby, 1956).

The differences between the 24 methods were the type of curve(s), the dependent variable of the relation(s), and the method for computing or positioning the curve(s). The type of curves included a single curve of suspended-sediment concentration (or discharge) versus water discharge, two seasonal curves--one each for the dormant season (October 1 to March 31) and growing season (April 1 to September 30), and three separate particle-size curves--one each for sand, silt, and clay. Particle-size curves also were prepared for the two seasons. For each type of curve, both suspended-sediment concentration and discharge were used as the dependent variable. Finally, three different methods were used to compute or position the curve. These included a linear regression, nonlinear regression, and a hand-drawn curve through group averages. The linear and nonlinear regressions were represented by equations 1 and 2, respectively:

$$\log S = \log a + b \log Q \quad (1)$$

$$\log S = a (1 - bQ^{-c}) \quad (2)$$

where S = sediment discharge, in tons or
sediment concentration, in mg/L,

Q = water discharge, in ft³/s, and

a , b , and c = regression coefficients.

The plotting positions of the group averages were the arithmetic average of suspended-sediment concentrations (or discharge) of all samples falling within a specified water-discharge range and the midpoint of the range.

Only samples collected for particle-size determinations for the Schuylkill River at Manayunk were used to compare the different methods. These data and the particle-size data for the other streams used in the evaluation are summarized in table 2. The variation in the relations between sand, silt, and clay and water discharge for the Schuylkill River is shown in figure 2. The variation is typical of the relation between instantaneous sediment concentration and water discharge measured during both the rising and falling stages. Usually the concentration of sediment for a given flow is higher during a rising stage than it is during a falling stage. The correlation coefficient and

standard error of estimate of the relations between the logarithms of water discharge and the logarithms of concentrations of sediment, sand, silt, and clay are shown in table 3. Generally, the relations between sediment and clay and water discharge have a lower standard error than the relations between sand and silt and water discharge. The growing season relations are as good or better than either the dormant or total relations.

Table 2. Summary of instantaneous water discharge and suspended-sediment concentrations for particle-size determinations and daily mean water discharges and concentrations during the indicated periods

[Water discharge is in cubic feet per second and suspended-sediment concentration is in milligrams per liter]

Station number	Period of particle-size and daily sediment sampling	Number of particle-size samples	Water discharge				Suspended-sediment concentrations			
			Instantaneous		Daily mean		Instantaneous		Daily mean	
			Minimum	Maximum	Minimum	Maximum	Minimum	Maximum	Minimum	Maximum
01473800	Oct. 1, 1953, to Sept. 30, 1979	86	1,800	73,600	274	93,400	55	1,970	1	3,140
01481500	Oct. 1, 1964, to Sept. 30, 1980	32	886	10,900	60	14,300	222	1,530	1	1,700
01517000	Oct. 1, 1954, to Sept. 30, 1967	42	12	887	0	364	41	3,720	0	1,070
01567000	Oct. 1, 1951, to Sept. 30, 1972	71	3,140	67,500	221	165,000	52	883	0	1,130
01567500	Oct. 1, 1954, to Sept. 30, 1970	64	6.1	950	1.9	799	82	3,550	0	986
01570500	Oct. 1, 1963, to Sept. 30, 1968	21 ²	48,700	526,000	1,700	408,000	121	510	0	540
	Oct. 1, 1971, to Sept. 30, 1978 ¹									

¹ Data for the 1972 water year were excluded from the record because the high sediment discharge associated with Hurricane Agnes in June 1972 would bias the average annual sediment discharge for the period of record.

² Samples collected between April 8, 1958, and February 25, 1981.

Table 3. Correlation coefficients and standard errors of estimate of relations between the logarithms of water discharge and concentrations of sediment, sand, silt, and clay in suspension

		Correlation coefficient	Standard error of estimate, in log units
Growing season	sediment	0.85	0.22
	sand	.87	.31
	silt	.85	.26
	clay	.73	.26
Dormant season	sediment	.75	.26
	sand	.84	.31
	silt	.66	.32
	clay	.75	.26
Total	sediment	.79	.25
	sand	.85	.31
	silt	.74	.30
	clay	.74	.26

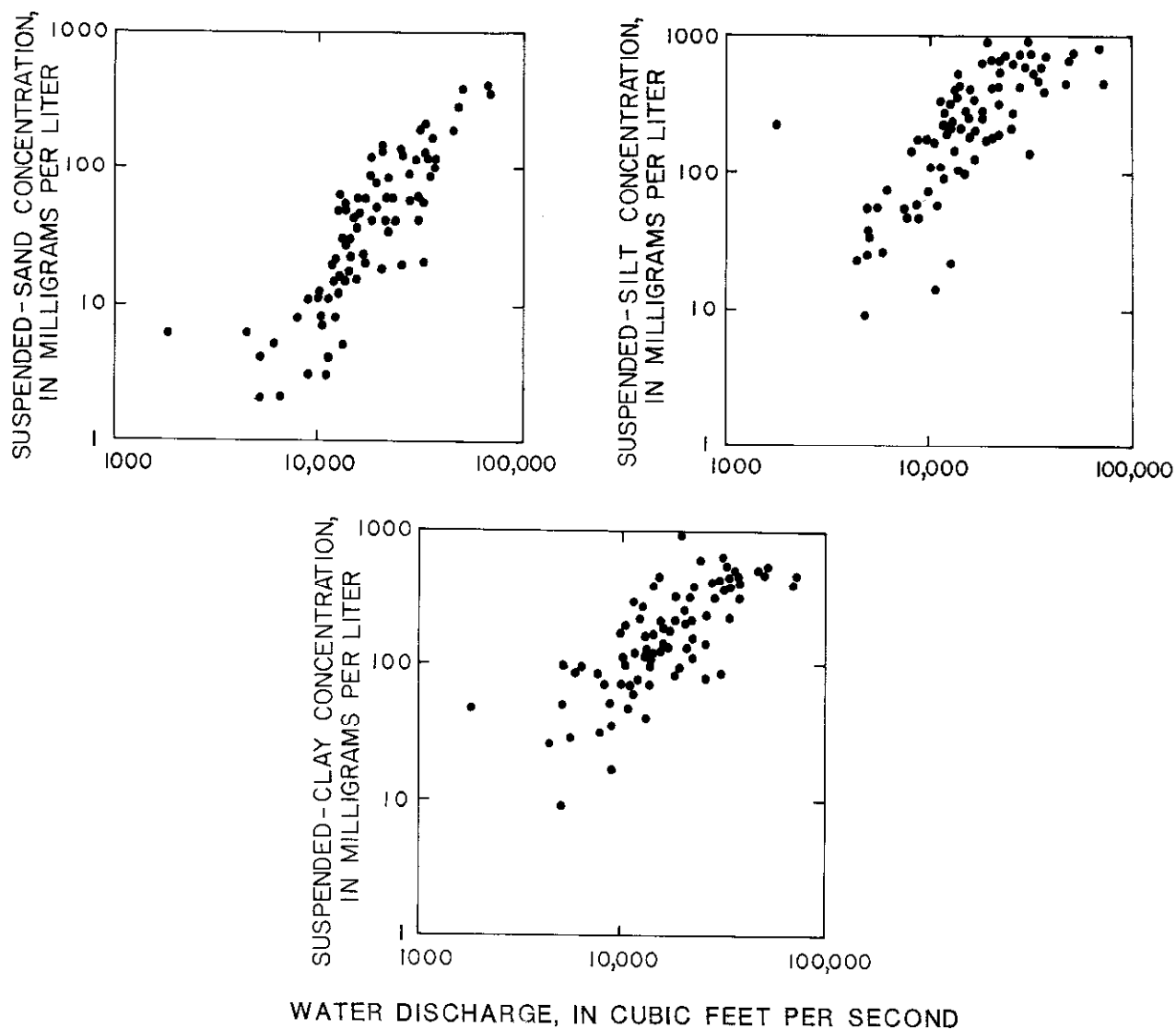


Fig. 2. Relations between concentrations of suspended sand, silt, and clay and instantaneous water discharge for the Schuylkill River at Manayunk, Pa.

ACCURACY OF THE DIFFERENT METHODS

The average annual suspended-sediment discharge estimated by 24 different methods ranged from 395,000 to 499,000 tons/yr (tons per year); the measured average discharge was 382,000 tons/yr (table 4). The mean discharge estimated by all the methods was 427,000 tons/yr and the standard deviation was 26,000 tons/yr. The means for the linear regressions, nonlinear regressions, and group average methods were 452,000, 418,000, and 412,000 tons/yr, respectively; the standard deviations were 26,000, 16,000, and 14,000 tons/yr, respectively. The average errors for the different methods were:

	Difference from measured sediment discharge, in percent
Linear regressions	18.4
Nonlinear regressions	9.3
Group averages	7.9
Concentration as dependent variable	11.3
Discharge as dependent variable	12.4
Single curve	13.1
Seasonal curves	11.4
Individual particle-size curves	12.0
Seasonal particle-size curves	11.0

Table 4. Measured and estimated average annual suspended-sediment discharges of the Schuylkill River at Manayunk, October 1, 1953, to September 30, 1979

	Sediment discharge, in tons/yr	Difference from measured discharge, in percent
Measured average annual sediment discharge	382,000	
Estimated average annual sediment discharges:		
LINEAR REGRESSIONS - Concentration data		
Single curve	480,000	25.7
Seasonal curves	444,000	16.2
Individual particle-size curves	424,000	11.0
Seasonal particle-size curves	424,000	11.0
LINEAR REGRESSIONS - Sediment-discharge data		
Single curve	445,000	16.5
Seasonal curves	442,000	15.7
Individual particle-size curves	462,000	20.9
Seasonal particle-size curves	499,000	30.6
NONLINEAR REGRESSIONS - Concentration data		
Single curve	426,000	11.5
Seasonal curves	432,000	13.1
Individual particle-size curves	435,000	13.9
Seasonal particle-size curves	418,000	9.4
NONLINEAR REGRESSIONS - Sediment-discharge data		
Single curve	403,000	5.5
Seasonal curve	431,000	12.8
Individual particle-size curves	397,000	3.9
Seasonal particle-size curves	398,000	4.2
GROUP AVERAGE CURVES - Concentration data		
Single curve	407,000	6.5
Seasonal curves	395,000	3.4
Individual particle-size curves	418,000	9.4
Seasonal particle-size curves	401,000	5.0
GROUP AVERAGE CURVES - Sediment-discharge data		
Single curve	432,000	13.1
Seasonal curves	409,000	7.1
Individual particle-size curves	432,000	13.1
Seasonal particle-size curves	404,000	5.8

The positive bias of all the estimates probably is caused by the use of only samples collected for particle-size determinations. These samples were collected only during storms when the concentration of suspended sediment was high in order to have sufficient material for laboratory analyses (Guy, 1969). In order to have a more normal distribution of samples, including both storm and baseflow samples, the data set for the Schuylkill River station was expanded to include water and sediment discharges for each October 15 and April 15 for the 26-year period of daily sediment records. This expanded data set was representative of a partial-record sediment station at which some baseflow and storm samples would be collected each year.

It was not possible to evaluate all 24 methods of computing sediment discharges with the expanded data set; however, three of the more suitable methods were tested. The three methods were (1) linear regression of suspended-sediment discharge and water discharge; (2) nonlinear regression of suspended-sediment discharge and water discharge; and (3) a group average curve of suspended-sediment concentration versus water discharge. These methods were tested because they are the traditional methods, they do not require particle-size data (which were not available for the October 15 and April 15 data), and they are the least complicated methods (annual regression requires one computation and seasonal particle-size regressions require six separate computations). The addition of October and April data resulted in a lower estimate of sediment discharge for each method that was tested (table 5). The linear regression and group average estimates were improved 41 and 64 percent, respectively, but the accuracy of the estimate based on the nonlinear regression was decreased about fourfold.

Table 5. Comparison of estimated average annual suspended-sediment discharge for the Schuylkill River at Manayunk with and without October 15 and April 15 data

	Sediment discharge, in tons/yr	Difference from measured discharge, in percent
Measured average annual sediment discharge	382,000	
Estimated average annual sediment discharges using annual relations between sediment discharge or concentration and water discharge:		
Linear regression (sediment discharge)		
--without October 15 and April 15 data	445,000	+16.5
--with October 15 and April 15 data	345,000	- 9.7
Nonlinear regression (sediment discharge)		
--without October 15 and April 15 data	403,000	+ 5.5
--with October 15 and April 15 data	307,000	-19.6
Group average (sediment concentration)		
--without October 15 and April 15 data	432,000	+13.1
--with October 15 and April 15 data	400,000	+ 4.7

The accuracy of the three methods used with the Schuylkill River data were tested with similar data collected at the other five streams and rivers. The particle-size data, summarized in table 2, were supplemented with daily mean data for each

October 15 and April 15 for the period of record. Table 6 shows the comparison between measured and estimated suspended-sediment discharge for all six streams used in the study. The linear regression model relating suspended-sediment discharge to water discharge produced reasonable results for each stream. Differences between measured and estimated discharges ranged from -7.0 to +30.8 percent. The estimate based on nonlinear regression between water discharge and suspended-sediment discharge for the Schuylkill River was 19.6 percent less than the measured discharge; the estimates were poorer for all the other stations. The only explanation for these differences is that the nonlinear model selected for the Schuylkill River data did not adequately represent the relations between suspended sediment and water discharge for the other sites. The group average method of defining the relation between sediment concentration and water discharge produced the best results for the Schuylkill River; it also produced an excellent estimate for the Susquehanna River (2.7 percent), but it produced the worst results for the other four streams. The differences between measured and estimated sediment discharges ranged from -17.6 to +132 percent.

Table 6. Comparison of measured and estimated average annual suspended-sediment discharge for six streams in Pennsylvania and Delaware

[Suspended-sediment discharge in tons per year]

Station	Measured	Linear regression estimate	Nonlinear regression estimate	Group average estimate
Schuylkill River at Manayunk, Pa.	382,000	345,000	307,000	400,000
Brandywine Creek at Wilmington, Del.	64,600	60,100	44,100	83,300
Elk Run at Mainesburg, Pa.	1,370	1,230	1,043	2,170
Juniata River at Newport, Pa.	291,000	381,000	390,000	522,000
Bixler Run near Loysville, Pa.	999	1,300	1,635	2,320
Susquehanna River at Harrisburg, Pa.	2,630,000	2,700,000	1,190,000	2,700,000

The scatter plots of suspended-sediment concentration versus water discharge for each of the six streams (fig. 3) provides some insight for explaining the wide variation in the differences between measured and estimated sediment discharges. The plots for the Schuylkill and Susquehanna Rivers exhibit reasonable straight line relations with a minimal amount of scatter; the estimates of long-term sediment discharge for these two sites were the best of the six test sites. The plots for the other streams indicate that much poorer relations exist between suspended-sediment concentration and water discharge. The plot for the Juniata River forms a straight line, but the scatter is much greater than either the Schuylkill or Susquehanna Rivers. The scatter is particularly significant for high water discharges. This could be caused by a large time lag between concentration and water discharge peaks or multiple concentration peaks. The

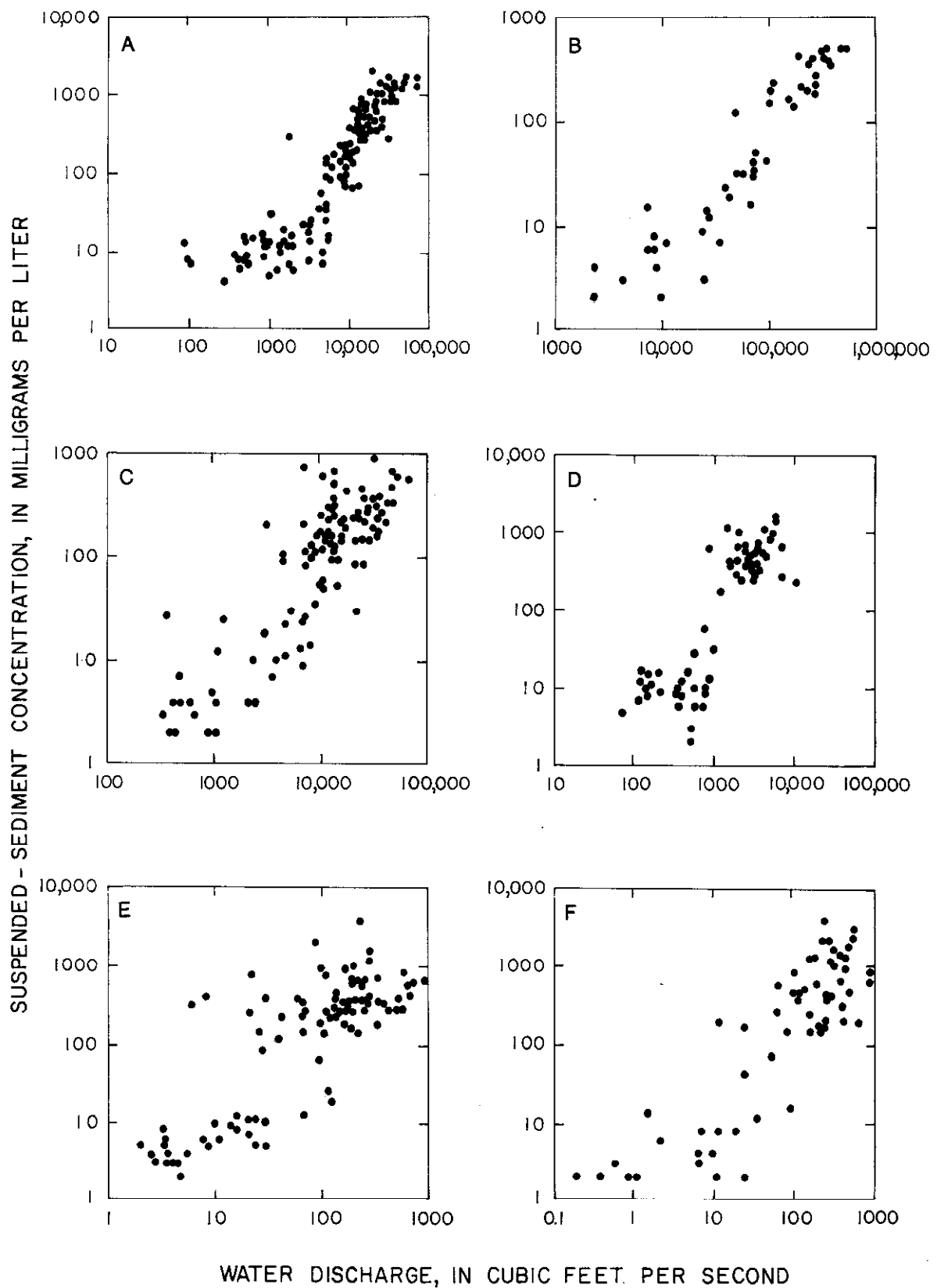


Fig. 3.--Scatter plots of suspended-sediment concentration versus water discharge. A, Schuylkill River at Manayunk, Pa.; B, Susquehanna River at Harrisburg, Pa.; C, Juniata River at Newport, Pa.; D, Brandywine Creek at Wilmington, Del.; E, Bixler Run near Loysville, Pa.; F, Elk Run near Mainesburg, Pa.

plot for Brandywine Creek shows a poor distribution of samples. Samples would need to be collected over a wider range of flow conditions before accurate estimates of long-term sediment discharges can be made. The plots for both Bixler Run and Elk Run show poorly defined relations throughout the range of water discharge. These plots probably are typical of small streams in which the sediment concentration at any given time is more a function of the amount of sediment available for transport and the intensity of rainfall and runoff rather than the total amount of runoff. None of the methods used to estimate long-term sediment discharge may produce reasonable estimates from instantaneous sediment concentration and water discharge data for small streams.

SUMMARY AND CONCLUSIONS

The extensive analysis of the sediment-transport and flow-duration curve method of estimating long-term sediment discharges indicated that many different methods provided accurate estimates for the Schuylkill River at Manayunk. Of the 24 methods used in the analysis, 21 of them provided estimates within 20 percent of the long-term measured discharge. Ten methods produced estimates within 10 percent of the measured discharge. The average difference between measured and estimated discharges was less for seasonal and individual particle-size curves than for a single curve; it also was less for relations between suspended-sediment concentration and water discharge than relations between suspended-sediment discharge and water discharge.

The use of only samples collected for particle-size determinations caused a positive bias of all the estimates. The addition of daily sediment and water discharge data for each October 15 and April 15 for the period of record resulted in lower estimates of suspended-sediment discharge. Differences between estimated and measured sediment discharges for the linear and nonlinear regressions and group average methods were -9.7, -19.6, and +4.7 percent, respectively.

The results of the Schuylkill River analyses were not duplicated when three of the methods were tested with data from five other streams. The linear regressions produced the best estimates of long-term sediment discharge for the five streams; the errors ranged from -7.0 to +30.8 percent. The errors for the nonlinear regression and group average methods ranged from -63.7 to +34.0 percent and -17.7 to +132 percent, respectively. Both the linear regression and group average methods will produce reasonable estimates of long-term sediment discharges if a uniform distribution of instantaneous sediment-concentration samples are available for the analyses. The nonlinear regressions will produce good results if the appropriate model is selected. None of the methods will produce good results for small streams.

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LOCAL SCOUR AT DROP STRUCTURES

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ABSTRACT

The scour at the toe of a vertical wall and at the toe of a sloping sill were investigated experimentally and analytically. Approximate relations for predicting the ratio of the scour depth to the (energy) critical depth were obtained for the two geometries. For the vertical wall, the sediment scoured out left in suspension, and the parameters needed to describe the scour phenomenon were the ratio of the (energy) critical velocity to the fall velocity and the drop in water surface in ratio to the critical depth. For the sloping sill, which is the recommended geometry, the sediment scoured out left as bed load, and the parameters needed to describe the scour phenomenon were the critical depth/sediment size ratio and the ratio of the size of the riprap protecting the sill slope to the critical depth. A follow-up study of flow and scour characteristics for a vertical drop followed by an apron resulted in rules for determining the length of apron required and the range of tailwater for proper flow.

THE PROBLEM

In the design of a new bridge founded on erodible material, it will (almost) always be wisest and most economical in the long run to make it invulnerable to the largest flood to be truly expected (Laursen, 1970). The Federal Highway Administration would seem to have come to agree with this proposition since in their suggested procedures for encroachments on floodplains they state, "...it is assumed... that the bridge itself will not fail" (Corry, et al., 1980). It has been suggested that old bridges also should be examined as to their vulnerability, and if found to be vulnerable should be made less vulnerable insofar as it can be done economically (Laursen, 1984).

In assessing the vulnerability of new or old bridges, it quickly becomes evident that everything that needs to be known about hydrology, hydraulics, river behavior, and scour is not known -- or is not known adequately. When trying to modify a bridge to make it less vulnerable to floods, it also quickly becomes evident that everything that needs to be known is not known. A sill structure downstream of an existing bridge or culvert is a possible modification which especially comes to mind if the problem is one of stream degradation. However, the sill structure must

resist the design flood if it is to save the bridge. To resist the design flood, the sill structure must not be undermined by scour at the toe. The reason for the research reported herein is that an equation to predict scour at such sills did not seem to be available.

FLOW AND SCOUR PATTERNS EXPECTED AND OBSERVED

Two sill structure geometries were investigated initially: a vertical wall, and a 1 vertical to 4 horizontal sloping sill. (Laursen and Flick, 1983). Ehlers (1984) followed up with a master's thesis on straight drop structure with an apron.

The flow and scour patterns expected at the vertical wall are shown schematically in Fig. 1 in which the various terms are defined. The critical depth of flow which exists upstream from the overfall (unless the approach flow is supercritical) is descriptive of the scale either as the reference length, the unit discharge, or the specific energy of the flow. The flow increases in velocity and energy as it falls through the drop in the water surface elevation. In the pool of the scour hole the jet expands and is turned up by the bottom of the scour hole. The bed material is greatly disturbed by the jet impact, and particles are cast up into suspension. Sediment leaves the scour hole in suspension, and the downstream slope of the scour hole is at the angle of repose except that it builds up and slumps down cyclicly. Note that the scour depth is defined as being measured from the downstream water surface.

The sloping sill shown in Fig. 2 is covered with riprap which needs to be covered with mesh and anchored. The flow and scour patterns are similar to those of the vertical wall with certain important differences. For all but small drops in the water surface elevation, the flow as it enters the scour hole pool is the normal depth of flow. Therefore, the ratio of riprap size to critical depth is a parameter of importance instead of the ratio of the water surface drop to the critical depth. As the limit of scour is approached, the sediment leaves the scour hole as bed load. Therefore, the ratio the critical depth to the size of the sediment (as a substitute for the ratio the particle shear to the critical tractive force) is a parameter of importance instead of the ratio of the (energy) critical velocity to the fall velocity of the sediment.

THE EXPERIMENTS AND THE RESULTS

The experiments were conducted in a two-foot wide flume with sands shown in Fig. 3 having median diameters of 0.30, 0.66, 6.1, and 14.7 mm (0.012, 0.026, 0.24, and 0.58 inches). The discharge was measured with a V-notch weir and was varied to give critical depths between 0.02 and 0.2 feet. The vertical wall was simulated by wooden boxes that could be stacked to various heights at the entrance of the flume. The tailwater was set to

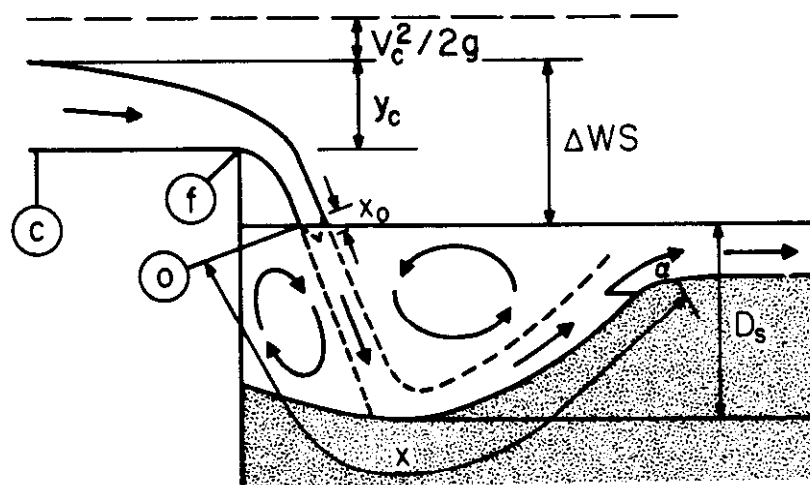


Fig. 1. Flow and Scour Pattern at a Vertical Wall.

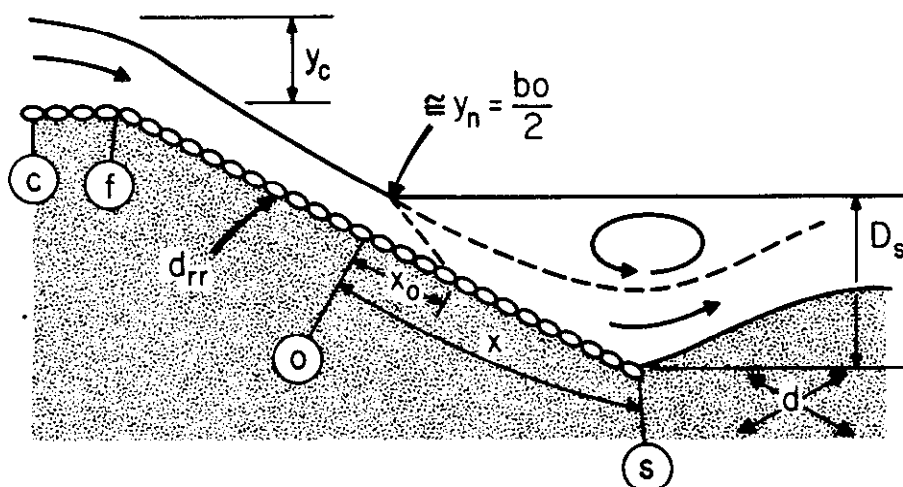


Fig. 2. Flow and Scour Patterns at a Sloping Sill.

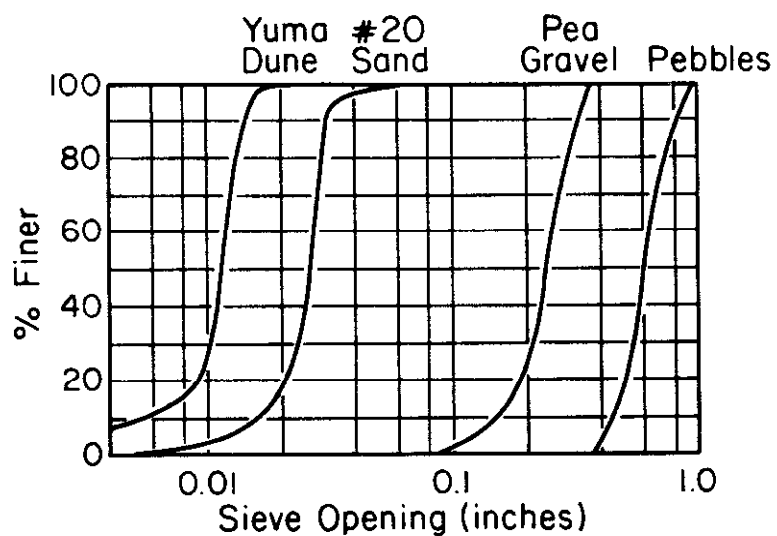


Fig. 3. Size Distribution of Sands and Gravels.

give values of $\Delta WS/y_c$ between 2 and 16. The sloping sill was also simulated with wooden boxes with tops having a slope of 1V:4H either bare (equivalent to a concrete surface) or covered with riprap which was pebbles (the coarsest of the sands used), or small rock of three-inch diameter. The discharges used gave critical depths between 0.06 and 0.4 feet, and the tailwater was varied to give values of $\Delta WS/y_c$ of 2 or 8. This latter parameter was of little consequence in the case of the sloping sill.

The experimental results for the vertical wall are presented in Fig. 4 and in Fig. 5 for the sloping sill. The curves in the two figures are based on very approximate analyses which were adjusted slightly in the coefficients and exponents. The final form of the prediction equation for the vertical wall is:

$$\frac{D_s}{y_c} = 8 \left(\frac{v_c}{w_o} \right)^{3/4} \frac{6 + \frac{v_c}{w_o}}{\sqrt{1 + \frac{2 \Delta WS}{y_c}}} \quad (1)$$

where in addition to the terms defined in Fig. 1, w_o is the fall velocity of a quartz sphere of the median diameter d of the stream bed material.

The final form of the prediction equation for the sloping sill is:

$$\frac{D_s}{y_c} = 4 \left(\frac{y_c}{d} \right)^{0.2} \frac{1}{3 \left(\frac{d_{rr}}{y_c} \right)^{0.1}} \quad (2)$$

where the terms are defined in Fig. 2.

A simple analysis of the size of riprap needed on the face of the sloping sill indicated that d_{rr} should be $2.8 y_c$. For this size of roughness, neither the evaluation of the critical tractive force, of the Manning n , or of the normal depth would be correct. Experimentally, it was found that the riprap was stable when the critical depth was less than 0.8 of the rock size. It is recommended that the riprap be covered with wire mesh and anchored; this will permit using smaller diameter rock and a lesser tonnage of rock.

In field situations where the critical depth can be two feet or more, and the depth of scour for the vertical wall can be well over ten times the critical depth, and for the sloping sill can be almost ten times the critical depth, consideration needs to be given to any effects that could lessen the scour. The prediction equations are for clear-water scour; if there is a sediment load of the size of the material being scoured out, the scour depth will be less because the boundary shear in the scour hole

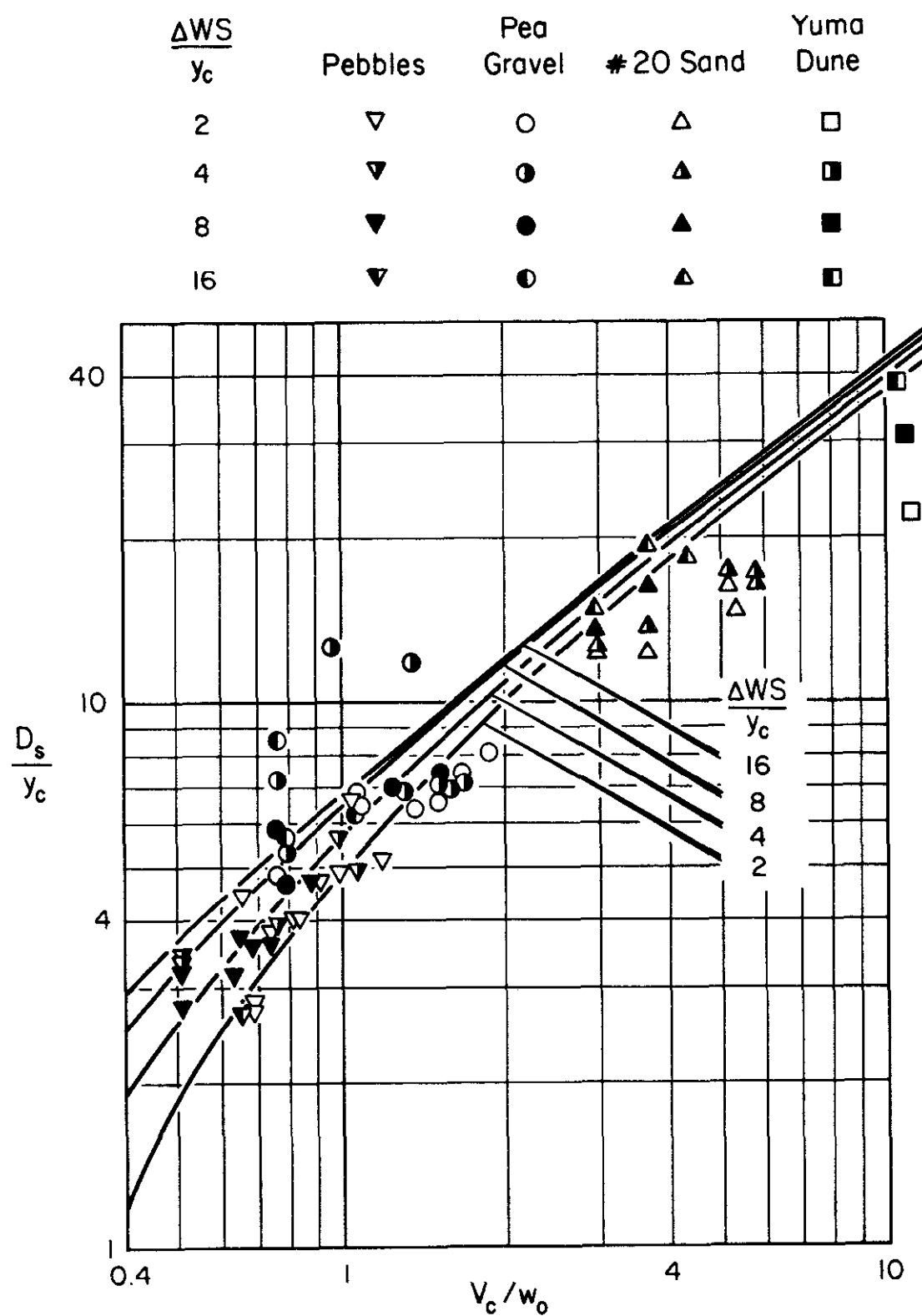


Fig. 4. Adjusted, Approximate Analytical Scour Relation for a Vertical Well.

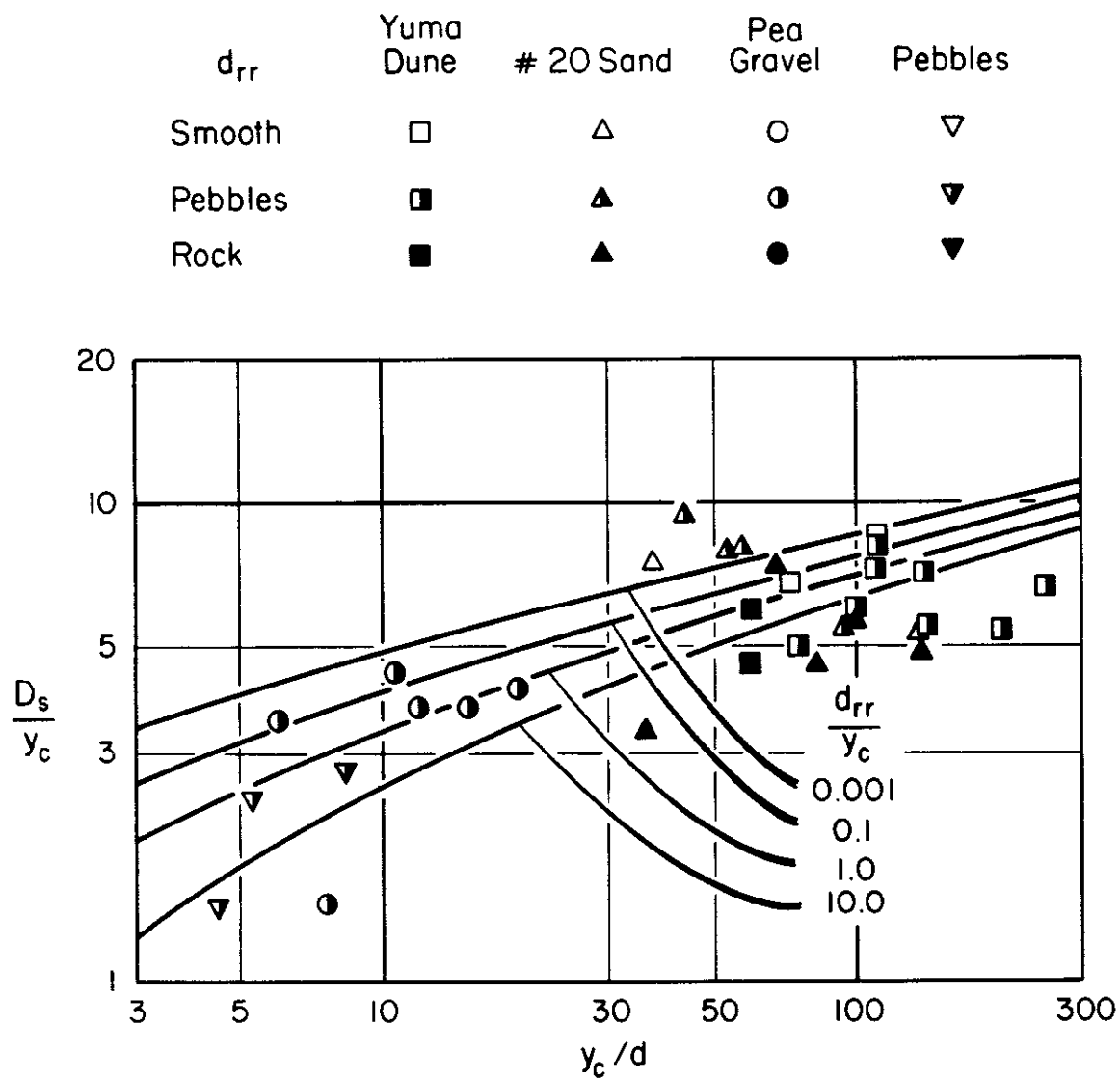


Fig. 5. Adjusted, Approximate Analytical Scour Relation for a Sloping Sill.

at the limit will be more (Laurson, 1952). A few runs were made with the sediment load which would be supplied by the flow at critical depth. In the case of the vertical wall, the depth of scour was three-fourths of what it would have been in clear-water flow. In the case of the sloping sill, the depth of scour was one-half of what it would have been in clear-water flow. In subcritical flow, the sediment supply to the scour hole would be less and the reduction in depth of scour would be less.

Another factor which could reduce the expected scour would be a quick rise in the hydrograph such that the limiting scour depth is not attained. In the experiments, most of the scour occurred in a very short time. Bridge pier scour in the field occurs so swiftly that the limiting state is sensibly always the scour experienced. This is probably not a factor to be counted on.

A third factor for reducing scour depth could be a change in sediment characteristics through self-armouring or deliberate riprapping. There will be a self-sorting action in the active scour process such that the bottom of the scour hole is composed of material coarser than the average size which is scoured out. This can be enhanced by preforming the scour hole and extracting the coarsest ten or twenty percent to be placed at the limiting scour depth. If the coarsest material thus available is not as large as desired, rock for riprap can be imported. The extra excavation, riprap rock, and replacement should be less costly than building the sill structure for the full depth of expected scour.

In the case of the vertical wall, the riprap modification was not entirely satisfactory. At the limit, the scour at the toe of the vertical wall was still very active with material being lifted into suspension, falling out on the downstream slope of the scour hole and then slumping back to the bottom of the scour hole. A very thick riprap blanket would be required to protect the underlying, finer material. The action could be reduced to almost nothing by increasing the scour depth to fifty percent more than would be predicted by Eq. (1).

In the case of the sloping sill, the riprap modification was very satisfactory. At the limit, scour sensibly ceases as the downstream slope of the scour hole flattens to the angle of repose. A nominal thickness of riprap should be sufficient with permeable fabric to prevent leaching of the underlying, finer material and a gravel layer to protect the fabric. With a "smooth" placed layer, the riprap was so stable that it could withstand an increase in the critical depth of 60%, or a doubling of the discharge per foot width.

It should be noted that a decrease in predicted scour on the basis of sediment-transporting flow, and a decrease in scour on the basis of riprapping cannot both be applied. A sediment supply of material finer than that which must be scoured out, the riprap, does not count -- it is now equivalent to clear-water flow.

THE STRAIGHT DROP STRUCTURE WITH AN APRON

Proper behavior of the straight drop structure with an apron requires that a jump form on, and be contained on, the apron. The geometry investigated is shown in Fig. 6 for both the submerged jet and hydraulic jump conditions. The structure was built of model gabions because the Arizona Department of Transportation ordinarily builds structures like this out of rock in wire mesh. Design of this drop structure involves deciding on the apron elevation and the apron length. The height of the drop will be primarily the amount of degradation being controlled.

The flow will behave properly only if the tailwater results in a flow depth at the end of the apron which is + 20% of the depth which gives a "good" hydraulic jump. If the tailwater is too high, the jump is submerged and the energy dissipation is poor with a wavy flow persisting downstream some distance. If the tailwater is too low, the jump is swept out and the stream bed is subjected to high shear. This means, of course, that the degradation must be predicted with precision and confidence. Even if the jump forms on the apron, the length of the apron must be sufficient to have a fully developed, normal flow coming off the end of the apron. The values needed for design can be obtained from Fig. 7.

The design values are plotted as a function of the drop number suggested by Bakhmeteff and Feodoroff (1943).

$$D_n = \frac{q^2}{gH^3} = \left(\frac{y_c}{H} \right)^3 \quad (3)$$

A trial-and-error solution is needed to obtain y_2 and H ; starting by assuming H is the amount of degradation. The length of apron should be at least $4y_2$ unless $L_d + 2y_c + H$ for the jump or $L_p + 2y_c + H$ for the submerged jet is larger.

RECOMMENDATIONS

The desired geometry of sill structure may be one of the three reported herein -- or some other shape for whatever the reason may be. If the geometry selected is close to one of these, it should be possible to estimate the scour to be expected; if not, it would be prudent to run some more experiments.

Based on the results of this investigation, the recommended geometry is the 1V:4H sloping sill for several reasons:

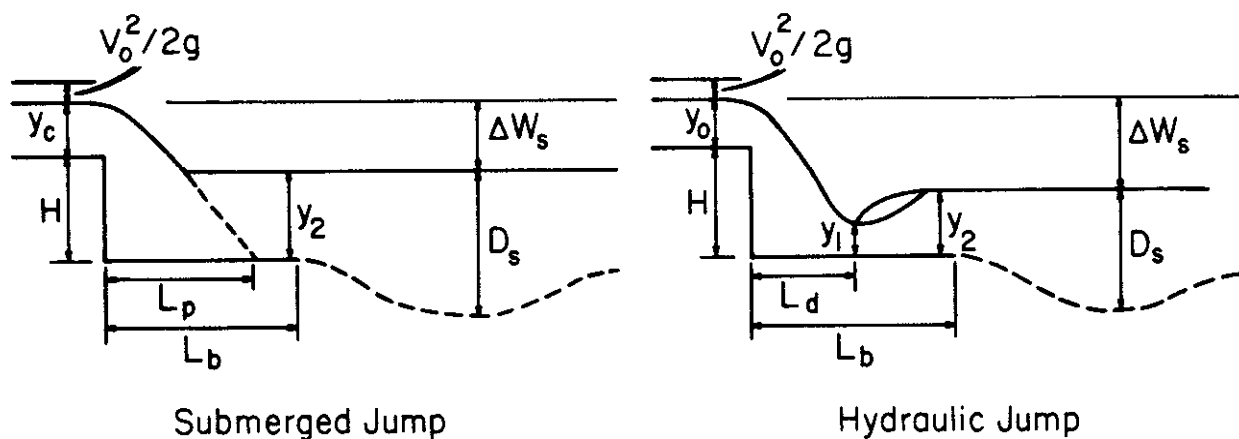


Fig. 6. Definitions for Straight Drop Structure.

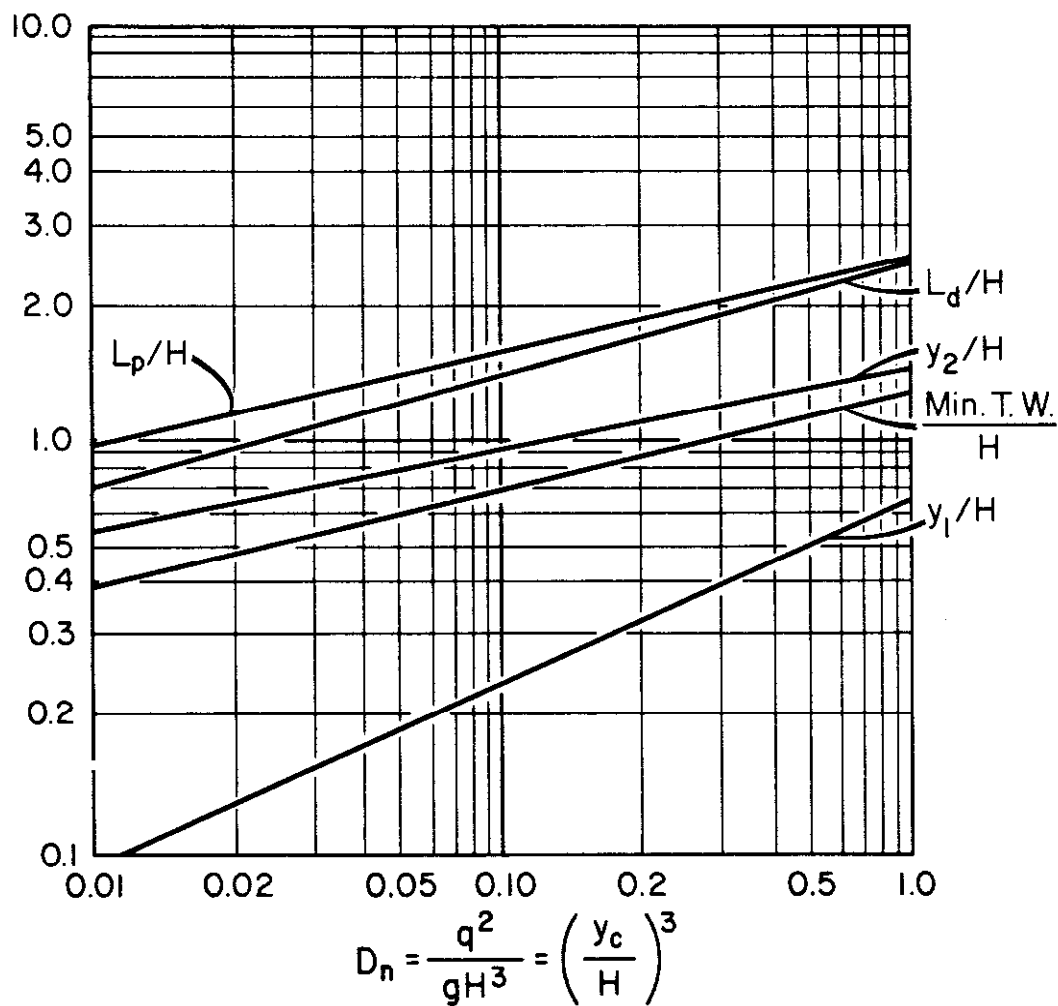


Fig. 7. Design Values for Straight Drop Structure.

1. Degradation in the future over the long run cannot be estimated with confidence. The sloping sill should be the easiest structure to extend downward at the toe if necessary. As the degradation exceeds the original estimate, the sloping sill will provide protection for all but the largest, rarest floods, giving time for repairs or modifications.
2. It should be possible to build the structure of the local stream bed material, imported riprap, wire mesh and anchors at a reasonably low cost.
3. Riprapping of the scour hole results in a very stable scour hole. Therefore, the predicted scour can be reduced, or a safety factor is built in. Note that the predicted scour should never be less than 50% greater than the normal depth downstream.

ACKNOWLEDGEMENTS

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SEDIMENT SOURCES AND SEDIMENT-TRANSPORT CURVES

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ABSTRACT

Studies of suspended-sediment transport in nine streams in northwestern California indicate that slopes and intercepts of suspended-sediment transport relationships are controlled by sediment sources and delivery processes. In terrain underlain by noncoherent rocks, sediment within channels is abundant and sediment transport relationships can be adequately described by single relationships which operate at relatively high levels. In terrain underlain by coherent rocks, sediment delivery to channels is more episodic and multiple relationships are required to adequately describe suspended-sediment transport relationships. Recognizing the interaction between the form of suspended-sediment transport relationships and dominant sediment sources aids in selecting the most appropriate relationship for a given water discharge and places such calculations on a sound basis with respect to drainage-basin process.

INTRODUCTION

The relationship between water and suspended-sediment yield, plotted as a suspended-sediment transport curve (SSTC), is commonly used to calculate suspended-sediment yield in streams. The form of these curves changes with space and time because suspended-sediment concentrations are affected by factors other than streamflow. The energy needed to transport available suspended sediment is usually a small percentage of total energy expended by a stream (Colby, 1963, p. 11). The quantity of suspended sediment transported by a given water discharge is affected by the availability of suspended sediment, derived either from within or from outside the stream channel. Sediment derived within the channel comes from mobilization of the streambed or from erosion of streambanks. Sediment derived outside the stream channel comes from fluvial hillslope erosion and mass movement.

Examples of how sediment yield is affected by drainage-basin processes are given by Colby (1956) and Galton (1984). Colby (1956) determined that suspended-sediment concentrations varied with time of year because storm intensities, soil conditions, vegetal cover, and snowmelt (all of which vary seasonally) affect sediment delivery from the land surface. In studying the magnitude and frequency of suspended-sediment transport in streams in the Lake Tahoe basin, Galton (1984) determined that the quantity of sediment transported by a given water discharge varied seasonally, reflecting variations in the quantity of snow cover and sources of runoff.

We provide additional examples of how the relationship between water discharge and suspended-sediment yield for selected streams (in northwestern California) is controlled by the availability of sediment and sediment-delivery processes. These factors control the slopes and intercepts of SSTCs. We also demonstrate the effects of sediment sources and sediment-delivery processes on the form of suspended-sediment transport curves for these streams in northwestern California. Although many of the effects of drainage-basin processes may be specific to the areas studied for this report, the recognition of dominant drainage-basin processes in other areas should facilitate interpretation of SSTCs.

DESCRIPTION OF STUDY AREA

Streams studied in northwestern California drain mountainous areas having a Mediterranean climate; however, rock types, hillslope steepness, and dominant mass movement processes differ among drainage basins. The area is characterized by two distinct terrain types. In one of these, drainage basins are underlain by noncoherent sandstones or pervasively sheared schist of the Franciscan Assemblage (Bailey and others, 1970) and are characterized by exceptionally unstable hillslopes. As a result, soil creep, earthflows, and debris sliding are active and deliver large quantities of sediment to channels. Because these processes are so active, stream channels commonly contain large quantities of stored sediment.

The other terrain type is underlain by relatively coherent rocks, has steeper hillslopes, and has mass movement dominated by episodic debris slides and debris avalanches, which primarily fail during major storms. This terrain is generally underlain by coherent sandstones of the Franciscan Assemblage. Mass movement in coherent terrain occurs more episodically than in noncoherent terrain. Throughout this paper these two terrain types will be referred to as noncoherent and coherent. Distinction between these two terrain types also has been recognized by Ellen and others (1982), who referred to coherent terrain as "hard terrain" and to noncoherent terrain as "soft terrain."

SUSPENDED-SEDIMENT TRANSPORT CURVES

Effects of contrasts in sediment storage and sediment-delivery processes on suspended-sediment transport curves were studied for nine streams (see figure 1). Sediment data for these streams, which drain either coherent or noncoherent terrain, were available for roughly the same years. To permit comparison among streams, data were normalized by dividing water discharge and sediment yield by drainage area. Sediment data for this study are in U.S. Geological Survey (1973-74, 1975-76), Iwatsubo and others (1975, 1976), or were obtained from Harvey Kelsey (Western Washington University, written commun., 1979).

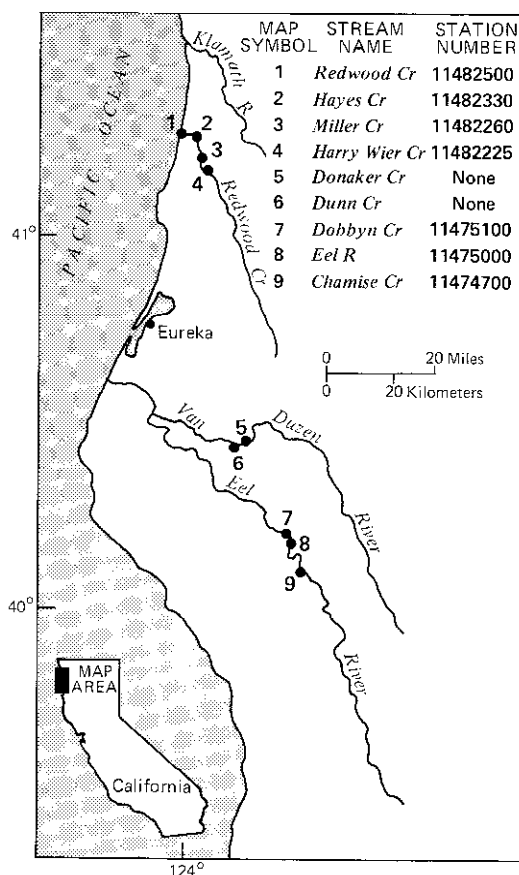


Fig. 1. Map showing location of study area.

Physical characteristics of streams studied are summarized in table 1. Because of the variety of rock types in the study area, most of the drainage basins are underlain by a combination of rock types. This is particularly true for larger drainage basins. Harry Wier, Miller, Hayes, Dunn, and Chamise Creeks represent streams dominated by processes associated with coherent terrain; and Redwood, Donaker, and Dobbyn Creeks, and the Eel River represent streams associated with noncoherent terrain.

Contrasts in Suspended-Sediment Transport Curves

Contrasts in dominant erosional processes in the two terrain types of northwestern California are manifested in contrasts in the slopes and intercepts of SSTCs. In noncoherent terrain SSTCs can be adequately defined by single relationships with generally high intercepts and low slopes. In contrast, at least two separate relationships are required to adequately describe SSTCs for streams draining basins containing large areas of coherent terrain. This is because there is a distinct change in slope at a water discharge of about $1 \times 10^{-1} \text{ (m}^3\text{/s)/km}^2$. For water discharges below this value, slopes of SSTCs are generally low; above this value the slopes are generally

Table 1.- Physical characteristics of study streams.
[Data compiled from Iwatsubo and others (1976), Nolan and others (1976), Kelsey (1980), Harden and others (1978), and Janda and Nolan (1979)]

Station name and No.	Drainage area (km ²)	Average ground slope (m/m)	Dominant rock type	Dominant hillslope process	Period of record used (water year)	Number of samples	Average annual suspended-sediment yield (Mg/km ²)
Harry Wier Creek near Orick 11482225	7.66	0.28	Coherent sandstone	Debris slides	1975-76	119	152
Miller Creek at mount near Orick 11482260	3.52	.30	do.	do.	1975-76	119	117
Hayes Creek near Orick 11482330	1.50	.32	do.	do.	1975-76	70	14
Redwood Creek at Orick 11482500	720	.26	Coherent and noncoherent sandstone	Debris slides, earthflow	1973-76	105	2,100
Donaker Creek near Bridgeville (no I.D. number)	1.17	na	Noncoherent sandstone	Earthflow	1975-76	18	24,199
Dunn Creek near Bridgeville (no I.D. number)	1.17	.29	Coherent sandstone	Debris avalanches, debris slides	1975-76	9	95
Eel River at Ft. Seward, 11475000	5,457	na	Coherent and noncoherent sandstone	Debris slides, earthflow	1973-76	111	1,356
Dobbyn Creek near Ft. Seward 11475100	159	na	Noncoherent sandstone	Earthflow	1973-76	70	3,000
Chamise Creek near Island Mountain 11474700	58	na	Coherent sandstone	Debris slides, debris avalanche	1973-76	52	1,607

high (see figure 2). Small water discharges in coherent terrain are usually associated with smaller values of sediment yield than are comparable values of water discharge in noncoherent terrain. However, slopes of upper limbs of SSTCs in coherent terrain are steep enough to cause this relationship to reverse at larger water discharges.

Analysis of Covariance

To test the statistical significance of differences in the slopes and intercepts of SSTCs for the two terrain types, SSTCs were grouped by terrain type and differences between these groups were tested using an analysis of covariance (Dixon and Massey, 1969, Chapter 12; Dixon 1983). This analysis tests for differences between regressions that describe SSTCs within groups by comparing slopes of individual regressions and mean values of suspended-

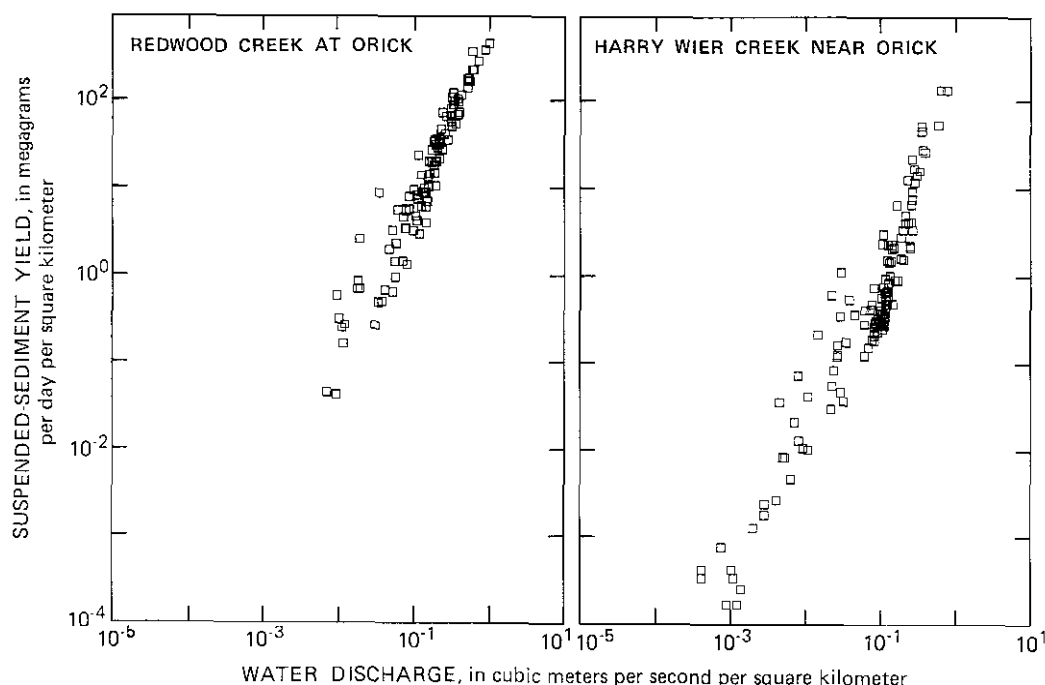


Fig. 2. Graph showing relationship between water discharge and suspended-sediment yield for Harry Wier Creek near Orick and Redwood Creek at Orick. Note increase in slope of the relationship for Harry Wier Creek above a water discharge of $1 \times 10^{-1} \text{ (m}^3/\text{s)/km}^2$.

sediment yield (after adjusting for differences in sampled ranges of water discharge). The significance of differences in slopes and adjusted mean suspended-sediment yield was tested against the F-distribution. According to the null hypotheses for these tests, the means and slopes are equal for both groups. The hypotheses were rejected if F-values were significant at the 95-percent confidence level. Results of the analysis of covariance are shown in table 2. Analysis was performed on logarithmically transformed data. For streams draining coherent terrain, only data from above the break in slope of the SSTC were used because most sediment transport and all channel-shaping events occur at water discharges above these breaks in slope.

Slopes and intercepts of SSTCs for streams dominated by processes associated with coherent terrain differs significantly from those for streams dominated by processes associated with noncoherent terrain (table 2). F-values for both tests are highly significant. Adjusted means of suspended-sediment yield are higher and slopes are lower for noncoherent terrain.

Table 2.- Results of analysis of covariance for study streams.
[Only upper limbs of SSTCs analyzed. Probability rounded to nearest percent]

Terrain group	Number of observations	Range in water discharge [$(\text{m}^3/\text{s})/\text{km}^2$]	Mean suspended-sediment yield ($(\text{Mg}/\text{d})/\text{km}^2$)	
			Group	Adjusted
Coherent	236	0.1-1.29	1.86	1.25
Noncoherent	304	0.008-1.94	12.0	16.4

Table 2.- Results of analysis of covariance for study streams--Continued

Terrain group	Slope	F-values			
		Equality of means		Equality of slopes	
		Value	Probability	Value	Probability
Coherent	2.34	686	0.00	16.1	0.00
Noncoherent	1.91				

Relationship between Drainage-Basin Processes and SSTCs

Contrasts in SSTCs from the two terrain types are apparently caused by contrasts in dominant sediment-delivery processes and by contrasts in the way sediment is stored and released from stream channels. In terrain dominated by processes associated with noncoherent rocks, mass movement is persistently active and delivers large quantities of sediment on an annual basis (Kelsey, 1980; Nolan and Janda, 1979). In addition, the inherently unstable streamside hillslopes in these areas fail easily during major storms (Harden and others, 1978). Large quantities of sediment delivered to channels during major storms can be sufficient to overload streams for decades (Nolan and Janda, 1979; Madej, 1984). As a result of storm effects and the persistent nature of mass movement, stream channels in basins underlain by noncoherent rocks commonly contain large quantities of sediment. Because source rocks are pervasively sheared and fractured, alluvium generally is small (sand and gravel) and readily transported by moderate flows.

This large supply of readily transportable sediment seems to be responsible for the fact that SSTCs from noncoherent terrain can be adequately described by a single relationship. Sediment supply is constant and abundant throughout the range of water discharge. As a result, SSTCs can be described by a single

relationship for each stream and the intercept, or level, of that relationship is relatively high (fig. 2). As water discharge increases, stream power and turbulence increase, and increasing quantities of sediment are removed from the channel.

In contrast with those in noncoherent terrain, stream channels in terrain dominated by processes associated with more coherent rocks contain a lesser volume of stored sediment, but that sediment is generally coarser. Large influxes of sediment to channels in coherent terrain are more dependent on extreme storms because such storms are necessary to trigger the debris avalanches and debris slides in these areas. Once introduced, storm-related sediment is flushed through these channels relatively quickly (see Pitlick, 1982).

The lack of readily transportable sediment at low flow in stream channels draining coherent terrain is reflected in SSTCs. Lower limbs of SSTCs apparently reflect the low availability of sediment during low flow; upper limbs reflect an influx of sediment, commonly at a water discharge of 1×10^{-1} (m^3/s)/ km^2 . Available bedload data indicate that this influx of additional sediment may be released from the streambed following the beginning of bedload transport. Figure 3 illustrates that bedload transport in Harry Wier Creek begins at about 1×10^{-1} (m^3/s)/ km^2 . Bedload movement possibly releases an additional supply of fine sediment from the streambed, which causes an increase in suspended-sediment transport. Once bedload movement occurs, the relationship between suspended sediment and water discharge steepens and remains generally uniform throughout the range of data collection. Because a

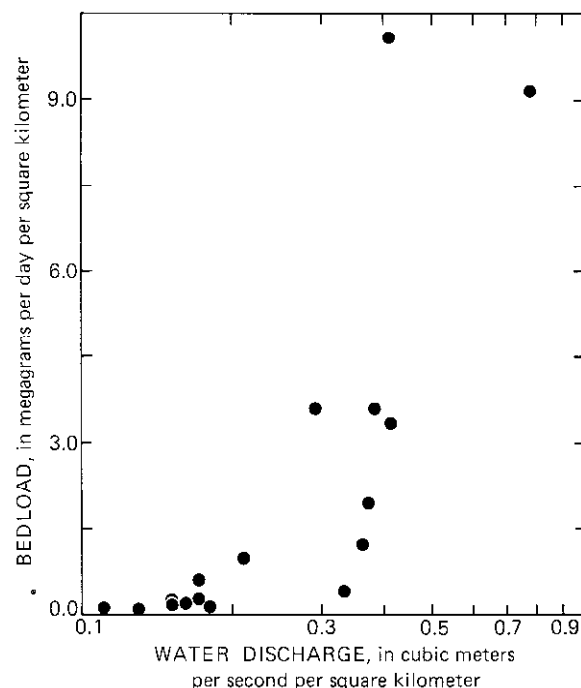


Fig. 3. Graph showing relationship between water discharge and bedload for Harry Wier Creek near Orick.

major storm did not occur during the study period, this study does not document effects of extreme events on SSTCs. Landsliding associated with such events could deliver sufficiently large quantities of sediment to cause additional changes in slope of SSTCs at very high flow.

Nolan and Janda (1981) suggested that the increase in SSTC slope also could be due to initiation of streambank erosion or removal of sediment from behind small unstable debris barriers within the channel. The available data do not rule out the effects of these two sources but indicate that bedload initiation must be considered a major cause of the change in slope of SSTCs.

When calculating suspended-sediment yield in coherent terrain, it is important to recognize the change in the relationship of suspended sediment to water discharge at about $1 \times 10^{-1} \text{ (m}^3\text{/s)/km}^2$. This is because a single SSTC based on all values would underestimate suspended-sediment yield for large values of water discharge (see fig. 2). Such a relationship would be affected by data associated with smaller water discharges and would have a significantly lower slope. Two relationships allow more accurate estimates of suspended-sediment yield throughout the full range of water discharge and avoid systematic underestimation of sediment yield during high flow.

SSTCs also provide information about the relative importance of various drainage-basin processes, and, therefore would be useful in sediment-budget studies which budget sediment yield by major contributing processes. Suspended-sediment-transport data may provide a means of estimating the relative importance of various sediment sources and delivery processes and thereby help focus work on the most important budget components. For example, SSTCs indicate that sediment-budget investigations in noncoherent terrain of northwestern California should focus on processes that provide a continuous supply of sediment (persistent landsliding and the long-term effects of major storms).

CONCLUSIONS

Because transport of suspended sediment is highly dependent upon sediment availability, suspended-sediment data provide substantial information about sediment sources and delivery processes. Understanding how these sources and processes affect sediment transport in a given drainage basin requires that sediment data be interpreted with a recognition of dominant drainage-basin processes. Recognizing the interaction between drainage-basin processes and sediment transport allows application of the most appropriate relationship of suspended sediment to a given water discharge. It also places sediment-yield calculations on a sound basis with respect to physical processes operating in a given drainage basin. The converse of this association also may be true. It may be possible to use suspended sediment to document the affects of individual sediment sources or delivery processes on suspended-sediment yield.

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MISSOURI RIVER SEDIMENT TRANSPORT RELATIONSHIP

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ABSTRACT

A comprehensive sediment and flow data collection program has been conducted at Yankton (River Mile 805.5), Sioux City (River Mile 752.3), Omaha (River Mile 615.9), and Nebraska City (River Mile 561.8) since the closure of the Gavins Point Dam in 1955. This is the most detailed set of field data for suspended sediment transport rate studies available at this time.

A sediment transport relationship was first developed based on these data. From this development, it was determined that (1) Toffaleti's flow velocity distribution was valid with the modification in which n (the exponent of the power law type of velocity distribution) is a function of sediment size, flow velocity and water temperature; (2) the vertical suspended sediment concentration profile in the entire flow depth can be treated as a single zone for the Missouri River rather than the three or four zones as proposed by Toffaleti; (3) a correction factor, C , needs to be multiplied to the well known Z value in Rouse's equation for vertical suspended sediment concentration distribution; and (4) a reference bed layer is needed to define the bed load sediment concentration. Unlike Einstein's assumption that the bed layer is equal to two times the sediment sizes, the thickness of bed layer is found to be a function of a Froude Number of the sediment particle size.

Another result of the study was the comparison of selected well-known sediment transport relationships. According to root mean square of the differences between measured and calculated values, the order of different models is (from best to worst): Shen and Hung (1971), Remodified Einstein method by Shen and Hung (1983), Modified Einstein's method by Colby and Hembree (1955), Yang (1973), Colby (1964), Toffaleti (1969) and Einstein (1950).

INTRODUCTION

Most research on sediment transport rates has been conducted in relatively small laboratory flumes, and it is difficult to extrapolate results for large rivers from this research. Many agencies have collected a great deal of field data from numerous rivers; but unfortunately, due to limited time and funds, insufficient comprehensive sediment data have been accurately collected in any single river to allow a reasonable analysis.

An additional difficulty is the lack of reliable field data in a zone close to the river bed. It is generally agreed that bed load cannot be accurately measured by a bed load sampler in large rivers with pronounced bed forms, and a suspended sampler cannot measure closer than 0.3 to 3 feet from the bed surface without disturbing the bed. The value of the ratio of the measured suspended sediment load to the total transport load is being actively debated. If estimates of the vertical flow distribution and the

suspended sediment concentration are made in the unmeasured zone and used to estimate the sediment load in the unmeasured zone, it will frequently be found that the measured load can be 10% to 120% of the total load. This ratio is a function of flow and sediment characteristics, and if it is not known, the true sediment transport rate of a river is not known and the accuracy of a transport equation cannot be determined.

In view of these difficulties, a carefully designed data collection program has been carried out at several stations on the Missouri River. The cross sections at each gaging station were divided into several verticals, and flow velocities as well as point suspended sediment samples were regularly measured at six points in each vertical. It is hoped that these vertical flow distributions and vertical suspended sediment transport rates can be established and extended to the river bed to determine the total sediment transport rate.

Detailed analysis of these data were presented by Shen and Wang (1979), and the purpose of this paper is to present a summary of these results.

MISSOURI RIVER DATA DESCRIPTIONS

Gaging Stations

Missouri River data have been collected by the Corps of Engineers and the U.S. Geological Survey at Yankton, Sioux City, Omaha and Nebraska City. (See Keown, Dardeau and Kennedy (1977) for detailed descriptions):

1. Water Surface Slope

Water surface profiles were measured from 1973 to 1976 during both high and low flow conditions. The water surface slopes were calculated by least-squared regression method. It was found that the surface slope at each station varied only slightly and without a definite trend. For convenience, the average water surface slope at each station was used for all flow conditions in the investigation. The average slopes were 0.000152, 0.000206, 0.000155 and 0.000208 for Yankton, Sioux City, Omaha, and Nebraska City, respectively.

2. Sediment Bed Materials

Sediment bed material samples were collected by a VSBM-54 sampler. Careful analysis indicated no definite relationship to season, water discharge, or temperature. Therefore, during analysis, the monthly average grain size distribution of the bed material was used for each station for flows occurring in that period. The overall average sediment size for the four stations was between 7.4 to 11.5×10^{-4} feet (or 0.23 to 0.35mm).

3. Point Flow Velocity

Point flow velocities were measured by a current meter, with 5 or 6 measurements usually made in each vertical.

4. Point Sediment Sample

The point sediment samples were collected by US P-61 or P-62 samplers. The quantity of each sediment sample collected was as much as five gallons. The size of most of the suspended sediment was between 1.7×10^{-4} and 9.7×10^{-4} feet.

5. Water Temperature

Water temperatures were recorded in degrees Fahrenheit at the beginning of sampling.

6. River Cross Section

Detailed survey of the channel cross section at the time of each data collection is not available; however, water depths were measured at 5 to 6 verticals across each cross section, and a rough approximation of the channel cross section can be obtained.

7. Range of Flow Conditions

The range of flow conditions used in this investigation is summarized in Table 1.

Table 1. Summary of Flow Conditions

	Yankton	Sioux City	Omaha	Nebraska City
Flow depth, ft	4 - 23	4.5 - 30	4.5 - 25	4.0 - 32
Flow velocity feet per second	1.5 - 5.5	1.0 - 7.5	1.5 - 8.5	1.5 - 9.5
Temperature, F	33-78	38-90	38-85	37-86
Flow discharge x100 cfs	137 - 348	156 - 463	111 - 485	198 - 718
Reynolds No., in 10	8 - 10	5 - 100	4 - 150	6.6 - 200
Froude number	0.07 - 0.26	0.1 - 0.32	0.1 - 0.4	0.07 - 0.4
Total data points available	123	238	329	325

DATA ANALYSIS

Since sediment transport rate is a product of flow velocity and sediment concentration, the accuracy of a sediment transport relationship is dependent upon the development of reliable expressions for these two factors.

Vertical Velocity Distribution

The two well-known methods to describe velocity distribution in turbulent boundary layer are (1) logarithmic velocity profile and (2) power function. Since Toffaleti (1969) found that for rivers, a power function agreed well with field data, this approach was used in this study. Toffaleti suggested that

$$\frac{u}{v} = n_1 \left(\frac{y}{D}\right)^{n_2} \quad (1)$$

where u is the velocity at a distance y from the river bottom, V is the average flow velocity, n_1 and n_2 are to be determined coefficients and D is the flow depth. The following theoretical relationship between n_1 and n_2 can be obtained by integrating both sides of equation (1) over the entire flow depth:

$$n_1 = 1 + n_2 \quad (2)$$

Actual velocity measurements collected at all four stations on the Missouri River indicated that

$$n_1 = 1 + 1.118 n_2 \quad (3)$$

Toffaletti (1969) found that n is only a function of flow temperature in the form

$$n_2 = 0.1198 + 0.00048T \quad (4)$$

where T is the water temperature in degrees Fahrenheit.

Based on Missouri River Data Collected, the following regression equation is obtained:

$$n_1 = 1.48d_{50}^{0.0226}V^{-0.0612}T^{-0.0223} \quad (5)$$

where d_{50} is the median sediment bed material size in mm and V is the flow velocity in feet per second.

This above relationship is quite different from equation (4) as determined by Toffaletti. Shen, Mellema and Harrison (1978) found that between summer and winter seasons, with the same total river discharge in this river reach, the bed form changes from dunes to plane bed, respectively. Flow velocity increases and roughness decreases. In other words, decreasing flow temperature and increasing flow velocity should decrease roughness and increase n_1 . Increasing d_{50} can probably decrease the bed form heights and thus decrease bed roughness. Actually, the phenomena is rather complex and the data may not be sufficiently accurate to determine these minor dependencies of n_1 values on d_{50} , V and T .

Vertical Suspended Sediment Concentration Distribution

Based on the assumption that upward sediment movements due to turbulent diffusion are balanced by the settling of sediment, Rouse (1937) developed the following equation:

$$\frac{c_y}{c_a} = \left(\frac{D-y}{y} \frac{a}{D-a} \right)^Z \quad (6)$$

where c_y and c_a are the suspended sediment concentrations at distances y and a respectively, from the river bed; D is the flow depth; and Z can be expressed by the following relationship:

$$Z = \frac{\omega}{\kappa U_*} \quad (7)$$

where ω is the sediment terminal fall velocity, κ is the Von-Karman Universal Constant and U_* is the shear velocity at the channel bottom. Einstein (1950) developed this well-known bed load function by assuming c_a in Equation (6) as the average sediment bed load concentration in a bed layer two sediment diameters deep, to obtain the total sediment transportation.

Suspended sediment concentration fluctuates a great deal and many suspended sediment samples are needed to establish a time average value. Since an insufficient number of samples are available, it was assumed in this study that the average value of many samples taken at different time periods under similar flow conditions can be used as representative values for that group. In each group of similar flow conditions the variations of flow velocity, flow depth and water temperature were less than 0.5 feet per second, 0.5 feet and 10°F respectively. Most of the concentration points were averages of three to five suspended sediment samples. Sediment concentration profiles of five size fractions were investigated: $1.7 - 2.4 \times 10^{-4}$ feet, $2.4 - 3.4 \times 10^{-4}$ feet, $3.40 - 4.89 \times 10^{-4}$ feet, $4.890 - 6.88 \times 10^{-4}$ feet, and $6.88 - 9.7 \times 10^{-4}$ feet.

The vertical suspended sediment concentration in the entire flow depth was assumed to be represented by Equation (6). A regression analysis was conducted on Missouri River data to determine the variability of Z value as defined by Equation (6). Z_t is defined as the corresponding Z value in Equation (6), and

$$Z_t = CZ \quad (8)$$

where the theoretical Z value is defined by Equation (7), Z_t is the exponent of the right hand side of Equation (6) as determined by Missouri River data, and C is a coefficient to relation Z and Z_t . Regression analysis indicated that:

$$C = 0.906 - 0.374 \frac{\omega^{0.471} n_1^{7.946}}{D^{0.41}} \quad (9)$$

where ω is the terminal sediment settling velocity in feet per second, n_1 is defined by Equation (1) and D is the flow depth in feet. As shown in Figure 1, the Missouri River data points are scattered considerably around equation (9), but a general trend exists. Of course, if the Z_t measured value is equal to the theoretical value of Z , then the value of C should be one. Missouri River data indicated that the values of C were about 0.75.

Sediment Concentration At A Reference Level

In order to obtain the suspended sediment concentration profile from Equation (6) and (7), the sediment concentration at a reference level must be known. Einstein (1950) assumed the bed load transport rate to be uniformly distributed in a bed layer with a thickness of two sediment diameters, and used this as the reference concentration in his bed load function. In this study, no bed load measurement was available in the Missouri River. Sediment concentration profiles in the unsampled zone were obtained by extending the vertical sediment concentration profile in the sampled zone down to the river bed.

A flow layer between the top of the river bed and 0.4 feet above the river bed was assumed to be the reference layer. Within this reference layer, the sediment concentration was taken to be uniform. It was found from the Missouri River data that the sediment load within the reference layer could be reasonably correlated to gd/V^2 , if the flow velocity V was taken to be average value across the entire flow depth and g was the gravitational constant.

If this reference layer is assumed to be less than 0.4 feet deep, the correlation between sediment load in the bed layer and flow drops off. If this reference layer is taken to be more than 0.4 feet deep, the correlation between sediment load in the bed layer and the flow does not improve significantly. A multiple regression analysis was used to obtain the following equation:

$$\Phi = \alpha \psi^{-1.088} \quad (10)$$

$$\text{where } \Phi = q_b / (\rho_s - 1)gd^3 \quad (11)$$

$$\alpha = S^{1.138} / (204178 \nu^{0.788} d^{1.06}) \quad (12)$$

$$\psi = \beta gd/V^2 \quad (13)$$

and q_b is the volume bed load discharge rate, ρ_s is the sediment density, S is the slope, ν is the kinematic viscosity, d is the grain diameter in feet, and β is a correction factor. β is a correction function of the kinematic viscosity of water and the shear velocity of the grain, u_*^1 . Denoting $\eta = 4.6 \nu^{0.33} / u_*^1$, β can be expressed as follows:

$$\begin{aligned} \beta &= 1 & \text{for } \eta \leq 1.2 \\ \beta &= 0.192\eta^{9.04} & \text{for } 1.2 \leq \eta \leq 1.3 \\ \beta &= \eta^{2.76} & \text{for } \eta \geq 1.3 \end{aligned} \quad (14)$$

Once the value of ϕ is calculated from Equation (10), the reference concentration c_a at $y = a$ (in which a equals two grain diameters) can be obtained by noting that

$$\begin{aligned}\phi &= \lambda \rho_f g \int_a^{0.4 \text{ ft.}} c dy \\ &= \lambda \rho_f g n_1 V \left(\frac{a}{D-a} \right)^Z c_a \int_a^{0.4 \text{ ft.}} \left(\frac{D}{y} - 1 \right)^Z \left(\frac{y}{D} \right)^{n_2} dy\end{aligned}\quad (15)$$

In this equation λ is a conversion factor from lb/s into tons/day, and Z is calculated from Equations (8) and (9). Transforming the integration variable y into $Y_* = y/D$, Equation (14) becomes

$$c_a = \phi / c' \left(\frac{a}{D-a} \right)^Z Y' \quad (16)$$

where

$$c' = \lambda \rho_f g n_1 V D \quad (17)$$

and

$$Y' = \int_{a/D}^{0.4/D} \left(\frac{1}{Y_*} - 1 \right)^Z Y_*^{n_2} dY_* \quad (18)$$

Y' can be numerically integrated using Simpson's Rule.

COMPARISON WITH OTHER SELECTED SEDIMENT TRANSPORT RELATIONSHIPS

Based on the computational purpose, the derivation of the sediment transport relationships can be classified into two types of procedures:

Type A: This type of procedure includes the methods developed to serve the design purpose, in which the sediment load is calculated directly from the flow condition and the bed material properties. Einstein's (1950), Colby's (1964), Toffaleti's (1969), Shen and Hung's (1971), and Yang's (1973) methods are in this category.

Type B: This type of procedure includes methods developed to calculate the sediment load in the unsampled zone from the measured sediment load. Among these are the Modified Einstein method by Colby and Hembree (1955), and the Remodified Einstein method by Shen and Hung (1983).

In this investigation, a Type A procedure was developed based on the Missouri River data. Since Toffaleti's method has been widely used in sediment load computation and has been considered as one of the most reliable, a comparison of the accuracy between it and the present method is made.

Since the actual total sediment loads in the Missouri River are not known, the comparison will be based on the RMS of the error of the calculated and sampled suspended loads in the sampled zone. In this comparison, the wash load, defined as the portion of the sediment load having sediment size less than d_{10} , is not included. The average water surface slope and the average grain size distribution of the bed material for all four stations are used. The RMS (root mean square) of error for all stations is summarized in Table 2.

Table 2. Comparison of Toffaleti's Method and the Present Method

Station	RMS of Error, tons/day/ft	
	Toffaletti's Method	Present Method
Yankton	13.5	7.5
Sioux City	43.8	37.8
Omaha	52.9	19.9
Nebraska City	62.4	25.3

Table 3. Comparison of Various Sediment Transport Formulae with Present Method - Omaha Station Data Only

Method Compared	RMS of Difference (tons/day/ft)
Shen and Hung (1971)	54.44
Modified Einstein Method (Shen and Hung, 1983)	65.53
Modified Einstein Method (Colby and Hembree, 1955)	75.55
Yang (1973)	88.74
Colby (1964)	104.01
Toffaletti (1969)	125.46
Einstein (1950)	562.99

The present method produces a smaller RMS error than does Toffaleti's method.

Comparisons of the present method with Einstein's (1950) method, the Modified Einstein method (Colby and Hembree, 1955), Colby's (1964) method, Toffaleti's (1969) method, Shen and Hung (1971), Yang (1973), and the Remodified Einstein method (Shen and Hung, 1983), were also done for the data collected at the Omaha station and the results are given in Table 3. Most interesting is that a simple regression equation by Shen and Hung (1971) shows the best agreement to the present method.

SUMMARY AND CONCLUSIONS

A sediment transport relationship is developed to describe the variation of total sediment transport rates with flow conditions in a reach of the Missouri River based on field data. Although this relationship should be able to predict the total sediment transport rate with reasonable accuracy, its ability to predict the transport rate of coarse material (sediment particles larger than 0.3 mm and moved in the unsampled zone near the bed) is not verified because the relationship is based only on suspended sediment concentration measurements in the sampled zone.

Specific conclusions are given in the following sections.

Vertical Velocity Distributions

A regression equation is developed to describe the variation of n_1 , the exponent of the power law type of velocity distribution, with flow condition. Missouri River data indicate that n_1 is a function of sediment size, flow velocity, and water temperature.

Vertical Suspended Sediment Concentration Distribution

The vertical suspended sediment concentration profile as developed by Rouse can be used for the entire flow depth. A correction factor of C is proposed by which to multiply the theoretical value of Z . From regression analysis, the correction C is found to be a function of fall velocity of the sediment particle, the flow depth, and n_1 , the exponent of the vertical velocity distribution.

Sediment Concentration In The Bed Layer

A reference bed layer is proposed to describe the sediment concentration near the bed. Within this layer, the sediment load is assumed to be uniform. It was found that the sediment load within the reference layer can be reasonably correlated to a Froude Number of the sediment particle size if the thickness of the reference layer is taken to be 0.4 feet above the bed. The magnitude of the sediment load in the bed layer is determined by extending the vertical suspended sediment distribution into this layer.

Comparison With Other Sediment Transport Relationships

1. In Sampled Zone Only

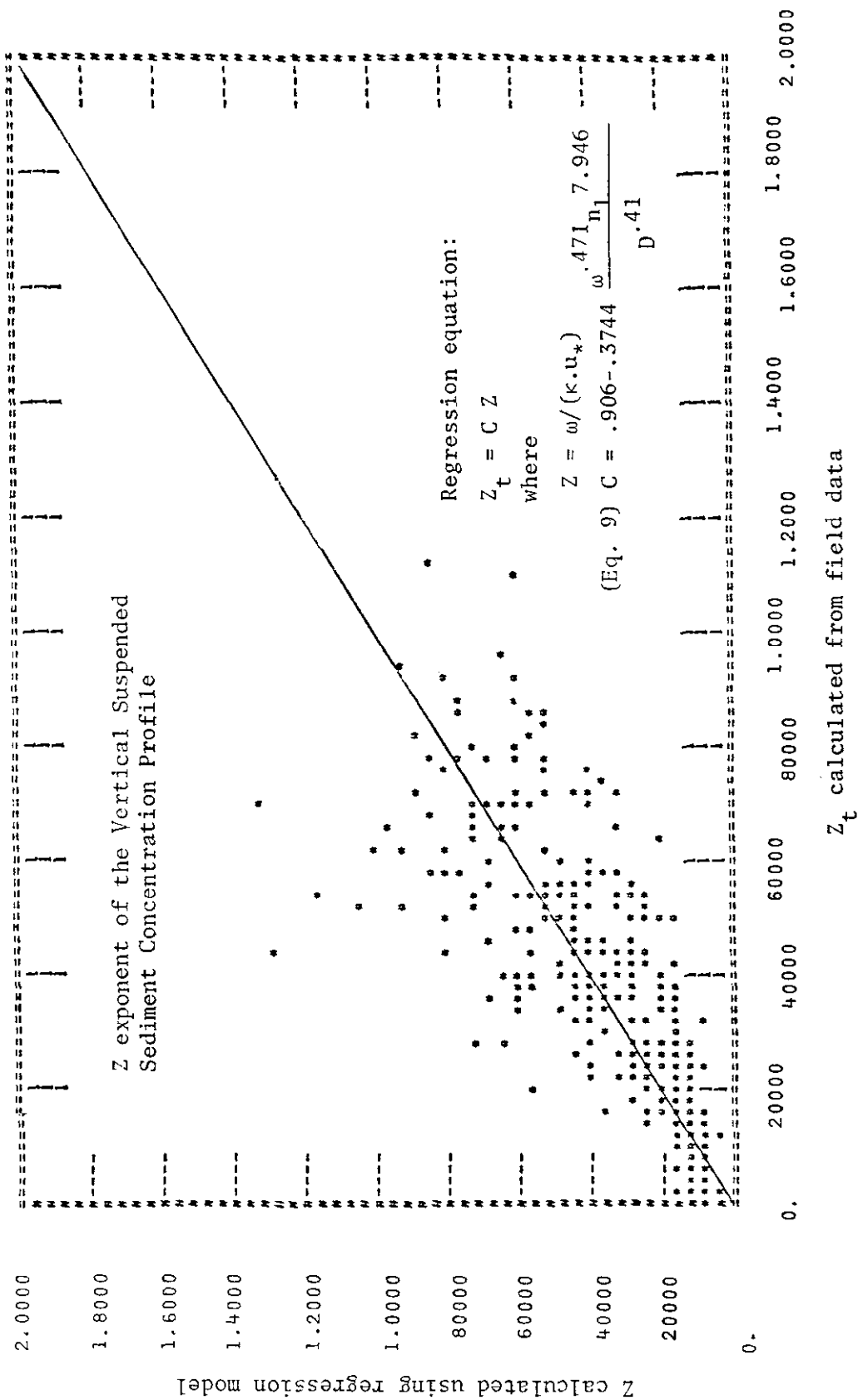
On the average, the RMS of error between our proposed method and the measured value in the sampled zone is about half of that between Toffaleti's method and the measured value.

2. Total Flow Depth with Other Methods(Omaha Data Only)

Based on Omaha Station data, results from the proposed method were compared with results by other methods. The other models are listed in order of the root mean square of the difference: Shen and Hung (1971), Remodified Einstein Method (Shen and Hung, 1983), Modified Einstein Method (Colby and Hembree, 1955), Yang (1973), Colby (1964), Toffaleti (1969), and Einstein (1950). In other words, a simple regression model based mainly on laboratory flume data as proposed by Shen and Hung (1971) gave the best results.

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Investigation of the Velocity Profile of the Missouri River

Figure 1. Variation of sediment concentration profile with flow condition.

DENUDATION RATES IN SELECTED DEBRIS-FLOW BASINS

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ABSTRACT

Denudation rates resulting from debris flows on Mount Shasta, California, on Mount Elbert in central Colorado, and in the Blue Ridge Mountains in Nelson County, Virginia, were determined by measuring volumes and ages of deposits. Results show that, where debris flows are frequent, denudation rates can be very high. If the debris is stored on lower slopes without entering the stream system, short-term sampling for sediment yield at streamflow gages sediment yields fails to reflect erosion in higher parts of the basin, and sediment-delivery ratios may approach zero. Data from the debris-flow areas provide extreme examples of sediment storage that may also be common in fully fluvial systems; they show that disequilibrium between hillslope and channel sediments may prevail over periods ranging from less than a decade to thousands of years.

INTRODUCTION

Denudation rates are commonly estimated by sampling streamflow for sediment and solute load and dividing by drainage-basin area. Erosion rates in upland areas, as measured by various techniques, however, are typically greater than fluvial transport rates would suggest. These observations indicate that (1) much eroded sediment does not leave many drainage basins but rather is stored on intermediate and lower slopes and as various forms of riverine deposition, and (2) geomorphically recent sediment-delivery ratios for many upland areas are significantly less than unity. The sediment-delivery problem has been recently summarized by Walling (1983).

Because field investigations and data-collection networks in undisturbed watersheds commonly define sediment yields but rarely include erosion data, there is a tendency to average yield data over an entire catchment and equate averages of yield and erosion. In doing so, recognition may be lost of locally high erosion rates and aggradation and sediment storage in certain parts of the basin. The difference between erosion and sediment storage may represent significant disequilibrium on and below hillslopes in steep mountainous regions.

Extreme to perhaps typical examples of the disparity between late-Holocene rates of erosion and sediment yield are provided by data from selected areas of documented debris-flow activity. These areas are glacial-meltwater basins on the flanks of Mount Shasta in northern California, the Halfmoon Creek basin in the mountains of central Colorado, and the Davis Creek basin in the Blue Ridge Mountains of central Virginia. These three areas represent diverse morphogenetic regions from mountains of the western, central, and eastern United States between 38 and 42 degrees north latitude.

Although areas subject to debris flows probably comprise a small part of the landscape, the examples illustrate the extent to which erosion or denudation can exceed sediment yield. In this study, only mechanical denudation processes are considered; chemical erosion is inferred to be a small part of total erosion, especially for the basins of California and Colorado.

Few published estimates of either denudation rates or sediment yields resulting from debris flows are available. The Mount Shasta data in this paper are reported by Osterkamp and others (1985), whereas the denudation estimates for the Halfmoon Creek basin and the Davis Creek basin (R. C. Kochel, Southern Illinois University, written commun., 1984) are previously unpublished. Estimates of sediment movement as debris avalanches, caused by Hurricane Camille in 1969, however, are available for basins adjacent to Davis Creek (Williams and Guy, 1973). The Jiangjia Gully in Yunnan Province, China, may have the highest reported erosion rate anywhere in the world (Kang and Zhang, 1984). During 6 years of investigation, 108 debris flows occurred. The basin consists of highly-sheared metamorphic bedrock and vegetation is virtually absent due to overgrazing by goats and deforestation from a near-by copper smelter. Denudation rates during the years 1965 to 1967 and 1973 to 1975 averaged about 60,800 mm/1,000 yr (millimeters per thousand years). A study of debris flows of April 1978, near Christchurch, New Zealand included estimates of erosion and deposition, but frequency and, therefore, denudation rates could not be determined. Material deposited during three days of flow activity, however, was estimated to be equivalent to the fluvial sediment yield of roughly 6,000 years in that part of South Island, New Zealand (Pierson, 1980). Pearce and O'Loughlin (1978) compiled erosion data for other New Zealand catchments in which mass-movement processes were presumed dominant. A literature summary of denudation rates by surface processes including landslides is presented by Saunders and Young (1983); the summary includes ground-loss data for mudslides and debris avalanches, but not for debris flows.

Denudation rates for debris-flow basins of Mount Shasta were determined by estimating the volumes of debris deposited by large, out-of-channel flow events. The smaller debris flows confined to channels were disregarded, as was fluvial transport of all erosion products. Rates of denudation were estimated by determining ages of debris flows, generally by dendrochronological techniques (Hupp, 1984; Osterkamp and Hupp, 1985), carbon-isotope analyses, and historical records. Denudation rates in the Halfmoon Creek basin of Colorado were estimated from volume measurements of debris-flow fans and levees that were also datable by dendrochronology. Volumes of the Davis Creek debris avalanches and flows may be considerably in error because of measurement difficulties; ages of the events were determined by carbon-isotope analyses of wood, charcoal, and buried soils (Kochel and others, 1982).

EROSION RATES IN DEBRIS-FLOW AREAS

Estimated rates of erosion, or denudation, in three catchments headed by glaciers on Mount Shasta (Fig. 1) are given in Table 1; geomorphic features, botanical data, and historical information indicate that the formation of debris flows from mass movements is the dominant erosion process in each of the basins. The rates of denudation are given as averages for the contributing subaerial watersheds although in all three cases only about 5 percent of the catchment has been subject to rapid erosion by stream incision and mass movement. Hence, the denudation rates of Mount Shasta basins (Table 1) are roughly 20 times greater when consideration is restricted to areas of active incision. Rates of denudation are commonly expressed in mm/1,000 yr; to permit comparisons, approximate conversions to Mg/km²/yr (megagrams, or metric tons, per square kilometer per year) are also given in Table 1. Sediment yields are typically given in Mg/km²/yr, but conversions are also provided.

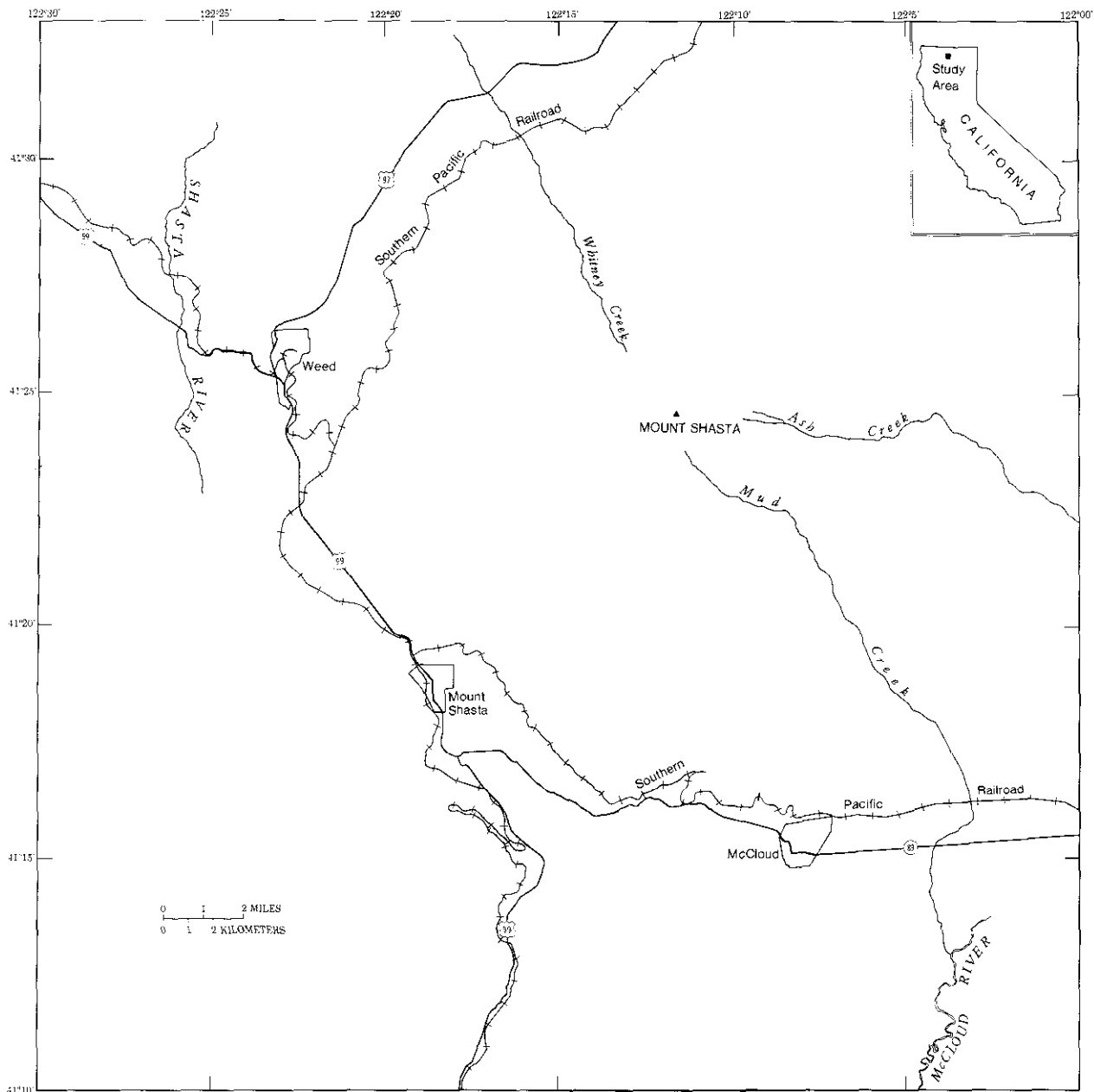


Figure 1. -- Map of Mount Shasta area, California, showing major streams and towns.

Estimated denudation for Whitney Creek may be the least accurate of the three Mount Shasta catchments. Although the Whitney Creek debris fan shows evidence of numerous late-Holocene depositional events, the only one for which a reasonably accurate volume is available is the most recent large flow of 1935 (Osterkamp and others, 1985). Erosion for the Whitney Creek basin, therefore, is based on a 50-year recurrence interval that may overestimate denudation and fan deposition (Table 1). Conversely, the denudation rate for Mud Creek is probably the most conservative and best documented. It is based on estimated volumes of five debris-flow episodes that range in age to about 1,200 years (Dickson and Crocker, 1953). The Ash Creek value is calculated from volumes of three flow episodes during the last 250 years. A fourth deposit about 800 years old yields an average denudation rate of about 3,000 mm/1,000 yr -- a rate regarded as unrealistically low owing to lack of data for the intervening 550 years.

Table 1. -- Denudation rates in selected debris-flow basins and comparative areas.

Basin or Location	Area (km ²)	Time (yrs)	Denudation (erosion)	
			(mm/1,000 yr)	(Mg/km ² /yr)
Whitney Creek, Mt. Shasta, California	10	50	11,000	30,000
Mud Creek, Mt. Shasta, California	16	1,200	4,000	11,000
Ash Creek, Mt. Shasta, California	12	250	6,400	17,000
South Halfmoon Creek tributary, Colorado	.39	46	260	700
Davis Creek tributary, Virginia <u>1</u> /	--	10,500	38	100
Davis Creek tributary, Virginia <u>1</u> /	--	6,340	24	65
Jiangjia Gully, Yunnan, China <u>2</u> /	29.2	6	60,800	162,000
Cropp River, New Zealand <u>3</u> /	28.5	12,000	13,000	34,400
Schesatobel, Austria <u>4</u> /	11.9	180	7,000	19,000
Los Angeles County, California <u>5</u> /	1.3	40	2,300	5,900
Ventura Avenue anticline, California <u>6</u> /	--	0.25-1.0x10 ⁶	600-2,500	1,600-6,400
Central Virginia <u>7</u> /	--	--	15-35	40-95

1. R. C. Kochel, written communication, 1984
2. Kang and Zhang, 1984
3. Griffiths and McSaveney, 1983
4. Eisbacher and Clague, 1984, p. 156
5. Scott and Williams, 1978, p. 34-35
6. Gilluly, 1949, p. 570-571
7. Judson and Ritter, 1964

With the possible exception of Whitney Creek, the denudation values for Mount Shasta basins (Table 1) are inferred to be underestimates because only large debris flows or observed deposition are considered. Sediment removed from the sides of Mount Shasta by fluvial transport, however, is minimal and does not appear to introduce significant error. Only small parts of large debris flows along Mud Creek reach the McCloud River on the south side of Mount Shasta (Wood, 1931), and all Holocene debris flows along Whitney and Ash Creeks have probably stalled before reaching perennial streams.

Considering that both volumes and frequencies of debris flows in the three Mount Shasta basins have large possible errors, the range of denudation rates suggested for the basins seems reasonable. The rates of 4,000 to over 10,000 mm/1,000 yr, as indicated by the several estimates, are inferred representative of longer-term averages for the upper slopes of Mount Shasta.

Relatively detailed volume measurements of fan and levee debris deposited by a tributary to South Halfmoon Creek, Colorado, between 1931 and 1977 suggest average erosion of about 260 mm/1,000 yr (Table 1). Steep mountain slopes in many parts of the Halfmoon Creek basin show relatively intense late-Holocene debris-flow activity, but most flow deposits have not been dated. Thus, the greatest error in the data from the South Halfmoon Creek tributary may be for the frequency or recurrence interval of debris flows. Botanical evidence of the recent flows as well as those of the last 185 years at the tributary site (C. R. Hupp, U.S. Geological Survey, written commun., 1984) suggests a frequency averaging one debris flow each 23 years. Flows of the South Halfmoon Creek tributary site were selected for study because virtually none of the debris entered the stream system except as washload immediately following deposition.

Davis Creek, Nelson County, Virginia, drains one of many small basins of the area that sustained catastrophic flooding and mass movements during Hurricane Camille in 1969 (Williams and Guy, 1973). From carbon-isotope dates of flood-plain and fan deposits below an upper-catchment site, Kochel and others (1982) found evidence of at least two other Holocene episodes of debris flows or debris avalanches. Volume determinations of the fan deposits by R. C. Kochel (written commun., 1984) led to evaluation of denudation rates for specific sites subject to recurring debris movement (Table 1). The rates, regarded as rough approximations, refer to mass movements only and may significantly underestimate actual rates owing to unknown losses of debris from the fan by fluvial erosion.

Estimated denudation caused by debris movement on Mount Shasta is more than one and two orders of magnitude greater than estimates for the South Halfmoon and Davis Creek tributaries, respectively. Several variables, including local relief, geology, hydrology, and vegetation, help account for this approximate 400-fold difference in erosion rates; the same variables provide explanations for the infrequent occurrences of mass movements at the Davis Creek tributary site in contrast to the relatively common debris flows in the Mount Shasta and South Halfmoon Creek drainage basins.

For comparison to the denudation rates of debris-flow areas considered here, erosion rates for a variety of other basins are provided in Table 1. Whereas the Jiangjia Gully represents highly disturbed conditions, the Cropp River basin of South Island, New Zealand is a relatively undisturbed catchment in which very high erosion rates throughout Holocene time appear to have resulted from the combined effects of high precipitation rates, rapid uplift, and soils easily mobilized as mass movements (Griffiths and McSaveney, 1983). The Schesatobel is a basin of the Austrian Alps that was denuded by lumbering in 1796; fluvial erosion and at least five debris flows since then have resulted in extensive denudation (Eisbacher and Clague, 1984, p. 156). In the Transverse Ranges near Los Angeles, California, erodible soils and fragile vegetation vulnerable to wildfire and human disturbance also cause high denudation rates (Scott and Williams, 1978). Long-term denudation, corresponding to rates of uplift, was estimated by Gilluly (1949) for the Ventura Avenue anticline, California, and Judson and Ritter (1964) estimated erosion rates due to fluvial processes for areas of central Virginia such as the Davis Creek basin.

SEDIMENT YIELDS

In contrast to the locally large denudation rates of debris-flow basins (Table 1), yields calculated from measurements of fluvial sediment transport in nearby streams are given in Table 2. Sediment yields for lands ranging from pristine forests to highly disturbed areas showing massive amounts of sediment movement are presented for comparison.

The Shasta River is the only through-flowing perennial stream of the Mount Shasta area for which fluvial sediment data are available. Whitney Creek is a tributary of the Shasta River although surface flow rarely continues its entire length without total infiltration. Periodic suspended-sediment samples collected during a 7-year period from the gage site Shasta River near Yreka, California, suggest a mean sediment yield of 11 Mg/km²/yr -- a rate less than 0.04 percent that of the estimated denudation rate for the upper Whitney Creek basin. Nine years of periodic suspended-sediment data from Halfmoon Creek near Malta, Colorado suggest

a sediment yield about 1 percent of the debris-flow denudation rate for the South Halfmoon Creek tributary site. Estimated denudation by mass movements at the Davis Creek tributary site is roughly ten times the corresponding sediment yield at the downstream gage site on the James River at Cartersville, Virginia (Table 2).

Comparative yields of fluvial sediment range from nearly 32,000 Mg/km²/yr in the Cropp River basin of New Zealand to less than 20 Mg/km²/yr for undisturbed forests of the United States (Table 2). Also provided for perspective are sediment yields for (1) a small basin (Stave Run) with active urbanization; (2) Levisa Fork, which drains a catchment with extensive surface mining; (3) two representative agricultural basins of central Kansas and central Maryland; (4) croplands of the United States in general; and (5) the Susquehanna River basin -- a large catchment of various land uses, including extensive surface mining. Relative to the comparison yields, data from streams near the debris-flow areas are suggestive of undisturbed, forested conditions (Table 2).

Table 2. -- Sediment yields near debris-flow sites and in other selected areas

Basin or Location	Area (km ²)	Time (yrs)	Sediment Yield	
			(Mg/km ² / yr)	(mm/1,000 yrs)
Shasta River near Yreka, CA 1/	2,060	7	11	4
Halfmoon Creek near Malta, CO 1/	61.1	9	7	3
James River at Cartersville, VA	16,206	18	<10	<4
Cropp River, New Zealand 2/	28.5	12,000	31,800	12,000
Stave Run near Reston, VA 1/	.21	2	3,100	1,080
Levisa Fork at Big Rock, VA 1/	769	9	580	215
Walnut River at Winfield, KS 3/	4,850	49	216	80
Monocacy River near Frederick, MD 4/	2,116	--	115	43
Average soil loss, U. S. cropland 5/	--	--	107	40
Susquehanna River at Harrisburg, PA 1/	62,419	13	48	18
Forested watersheds, eastern U. S. 6/	≤5	--	16	6
Forested watersheds, western U. S. 6/	≤5	--	16	6
Forested land, eastern U. S. 7/	--	--	11	4

1. From U. S. Geological Survey Water-Data Reports, various years
2. Griffiths and McSaveney, 1983
3. Osterkamp, 1977
4. Wolman, 1964
5. U. S. Soil Conservation Service, 1979
6. J. H. Patric, written communication, 1983
7. Patric, 1976

DISCUSSION AND IMPLICATIONS

The data in Tables 1 and 2 show that rates of sediment movement, as debris flows from upland basins, can be very high, comparable to the highest measured denudation rates by fluvial processes. As shown by the Davis Creek and other central Virginia data, the sediment moved during one mass-movement event may be equivalent to the denudation suggested by sediment yields for periods of hundreds to thousands of years.

Tables 1 and 2 also indicate that, for geomorphic and shorter time scales, natural sediment-delivery ratios can be significantly less than 100 percent. Gong and Xiong (1980) report an extreme of the sediment-delivery problem for the Yellow River basin in China, where delivery ratios approach 100 percent. Occurrences of loess deposits, active gully erosion, and persistent hyperconcentrated sediment flows keep extraordinarily large percentages of the fine-grained eroded sediment suspended and in transit. It is assumed that small, steep mountain basins subject to debris flows of coarse materials represent the other extreme in sediment-delivery ratios. In these environments, delivery ratios are very small -- less than 1 percent -- and they may approach zero. Debris flows of Mount Shasta and the South Halfmoon Creek tributary site provide dramatic examples of very low sediment-delivery ratios. In these cases, averages of hundreds to tens of thousands of $\text{Mg}/\text{km}^2/\text{yr}$ of sediment are transported from the areas of active alpine erosion and are stored downslope. Because of local conditions, the examples given here for Mount Shasta and the South Halfmoon Creek and Davis Creek tributary sites appear to represent sediment-delivery ratios that approach zero for small mountain basins subject to debris flows.

The very large disparity between erosion and sediment yields that is apparent for some debris-flow areas is an extreme example of late-Holocene geomorphic disequilibrium that has been identified by many investigators. The extent of the disparity in drainage systems subject only to fluvial erosion is rarely as great as it is at Mount Shasta or Halfmoon Creek, but where both erosion and sediment-yield data are available and reliable, delivery ratios commonly are less than unity. Particularly in relatively high-yield basins, such as many of the western United States (see, as examples, Emmett, 1974; Frickel and others, 1975; Hadley and Shown, 1976), storage of eroded sediment within the fluvial system can be substantial. Cases of sediment-delivery ratios approaching 100 percent in high-yield basins, such as those of the Yellow River in China and the Cropp River in New Zealand (Tables 1 and 2), may be exceptional.

After accounting for tectonic uplift, it seems clear that over geologic time scales sediment-delivery ratios should closely approximate 100 percent; Gilluly (1949) used this assumption in erosion estimates for the Ventura Avenue anticline (Table 1). Hillslope equilibrium, however, is event and time dependent, and the reasonable assumption of equilibrium through geologic time is not reasonable when applied to geomorphic or shorter time scales. The storage of late-Holocene debris flows and the resulting low values for sediment-delivery ratio provide emphasis for this conclusion. The most significant information gaps appear to be locations, residence times, and diagenetic alterations to sediments in storage in watersheds.

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MISSOURI RIVER AGGRADATION AND DEGRADATION TRENDS

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INTRODUCTION

The U.S. Army Corps of Engineers, Missouri River Division, has constructed six large multi-purpose dams in tandem along the upper Missouri River for flood control, power production, irrigation, water supply, augmentation of river discharges for navigation and water quality, recreation, and fish and wildlife benefits. These dams now intercept the sediment from one of the most sediment producing regions in the continental United States. Sediment measurement records indicate that an average of 135 million tons of sediment per year was transported by the river as suspended load prior to construction of the dams. Most of the sediments previously passing through the system are now deposited within the six-dam river-reservoir system. This has had a far-reaching influence, not only within the individual reservoir projects themselves, but also on the open river reaches between the projects and downstream from the lowest dam. Our experience to date and the impacts of the reservoir effects are discussed below.

MISSOURI RIVER AND RESERVOIR SYSTEM

The Missouri River mainstem reservoir system is composed of six large earth embankments that impound a series of lakes that extends for 1055 miles (Figure 1). Fort Peck Dam, the oldest of the six, was closed in 1937. Its embankment volume of 125.6 million cubic yards makes it the largest hydraulic fill earthen dam in the world. Fort Randall Dam was closed in 1952, followed by Garrison Dam in 1953, Gavins Point Dam in 1955, Oahe Dam in 1958 and Big Bend Dam in 1963. At their multiple purpose pools, the lakes provide nearly one million acres of water surface area and extend a total of 750 miles. Only 327 miles of open river remain between the lakes, although there are 805 miles of open river downstream from the lowest dam (Gavins Point) to the mouth of the Missouri River where it enters the Mississippi River at St. Louis, Missouri. The reservoirs contain a total of 75 million acre-feet of storage space, more than three times the average annual flow past Gavins Point Dam.

The Missouri River is formed in southwestern Montana in the high mountain area around Yellowstone National Park. The river flows through steep, narrow canyons, and is characterized by slopes of 0.00114 (6 ft/mile), until the river emerges onto the high plains area. The slope then flattens gradually to 0.00019 (1 ft/mile) at the approach to Fort Peck Reservoir and maintains approximately this slope with local variations all the way to the mouth. Characteristics of the open river reaches between the dams and their lakes are summarized in Table 1. Downstream from Gavins Point Dam the river remains in a semi-wild state for a distance of 59 miles in which it is free to meander throughout a wide flood plain. Downstream from this point the river is confined by revetment and dike structures into a single channel developed for bank stabilization and open river navigation, and varies in width from 700 ft

FIGURE 1
CORPS OF ENGINEERS
MISSOURI RIVER DIVISION

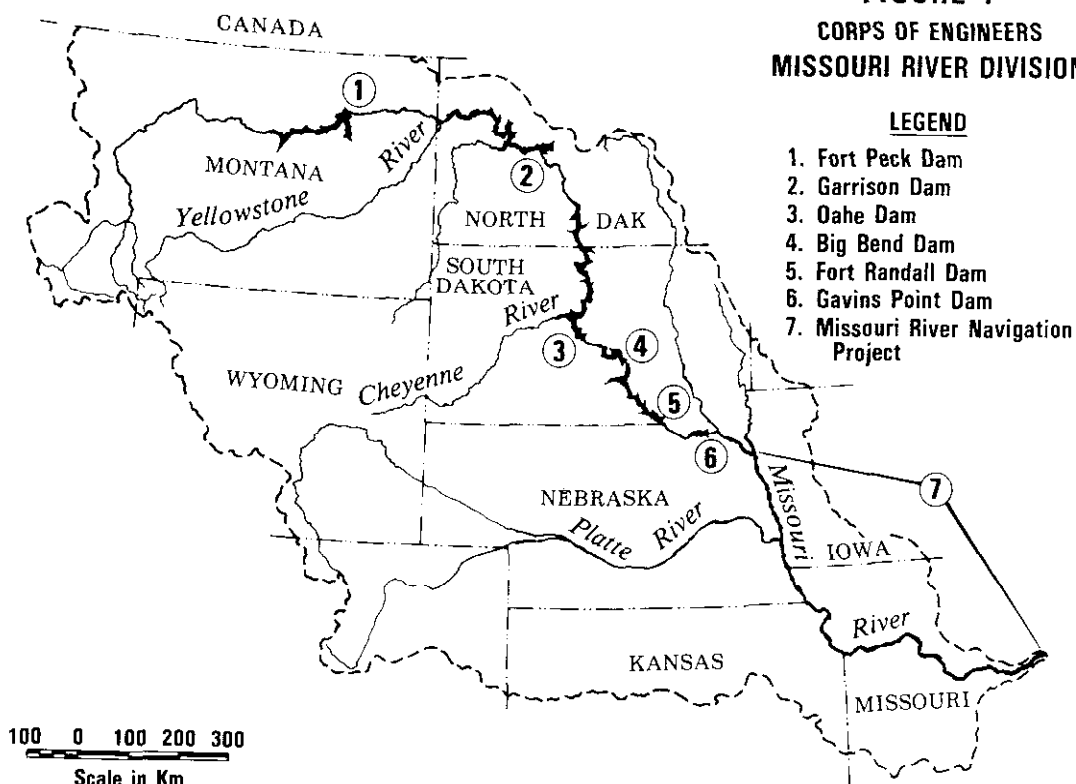


TABLE 1
OPEN RIVER REACHES GEOMETRY, MISSOURI RIVER

Reach	Open River Length (miles)	Valley Width (miles)	River Channel Width Between High Banks(ft)	Percent Valley Occupied by Channel
Fort Peck to Garrison	190	3	1,100	7
Garrison to Oahe	88	2	2,100	25
Oahe to Big Bend	5	0.6	1,000	30
Fort Randall to Gavins Point	44	1	2,100	30
Gavins Point to Ponca State Park	59	10	2,500	7

at Ponca, Nebraska, 600 ft at Sioux City, Iowa to 1100 ft at St. Louis, Missouri, a distance of 750 miles.

Prior to the reservoirs, the mean annual discharge was about 32,000 cfs at Sioux City, Iowa. Flood flows greater than 100,000 cfs and summer or winter sustained low flows less than 10,000 cfs were experienced nearly every year. Record floods and record low flows differed from the mean annual discharge by more than ten fold. Flow regulation by the reservoir system has profoundly changed the discharge regimen. Although the average annual discharge at Sioux City has not changed appreciably, the reservoir releases are limited in years of near-normal runoff to a high of about 35,000 cfs during the navigation season and lows of 14,000 to 21,000 cfs in the winter. Since full regulation began in 1954, the highest sustained release has been about 60,000 cfs and the lowest, about 10,000 cfs. The regulation effects on flood discharges diminish with distance downstream from Gavins Point Dam as the Missouri River picks up an additional 250,000 square miles of tributary drainage.

The reservoirs also have had a profound effect on downstream sediment loads. In the natural river the average annual sediment load at the site of Gavins Point Dam was about 135 million tons per year, although the loads for individual years varied from the average by a factor of two. The loads were composed of about 20% sandy bed material load and about 80% silty and clayey wash load. With dam closure, virtually all the incoming sediment was entrapped in the reservoirs and the loads just below the dam sites were reduced essentially to zero. Downstream from the dams with open river reaches, the river recovers its bed material from the sandy river bed after a few miles. The wash load of silts and clays, however, remains drastically reduced, although the river picks up some wash load from bank erosion and from downstream tributaries. Figure 2 shows the annual sediment loads and their grain size composition for the period from 1940 to 1977 at Omaha Nebraska, 195 miles downstream from Gavins Point Dam. The drastic change in sediment load characteristics coincides with Fort Randall and Garrison Dams going into full operation in 1954 and Gavins Point in 1956. Similar trends are prevalent in all the open river reaches within the reservoir system.

RESERVOIR DEPOSITION

Pre-construction estimates of sedimentation rates in the reservoirs were based on field sampling of the suspended load at selected locations along the upper Missouri River and principal tributaries. Most field information included no more than 10 years of data obtained mostly during periods of below-normal flow. Annual average sediment loads from the measured data were correlated with measured annual streamflow volumes and then extrapolated back over the entire period of streamflow measurement since 1898 to estimate long-term average annual sediment loads. A ten percent correction was added to account for the unmeasured sand load below the lowest sampling point. Using the unit weight of 75 pounds per cubic foot dry weight, the "Original Estimates" of sediment deposition rates were obtained. Since the original estimates were made, more data have become available. Actual reservoir survey data have shown that the sediment deposition density ranges from 60 to 75 pounds per cubic foot. These extended data have been the basis for the "Current Estimate" shown in Table 2.

FIGURE 2
MISSOURI RIVER
AVERAGE ANNUAL SUSPENDED SEDIMENT LOAD
@ OMAHA, NEBRASKA

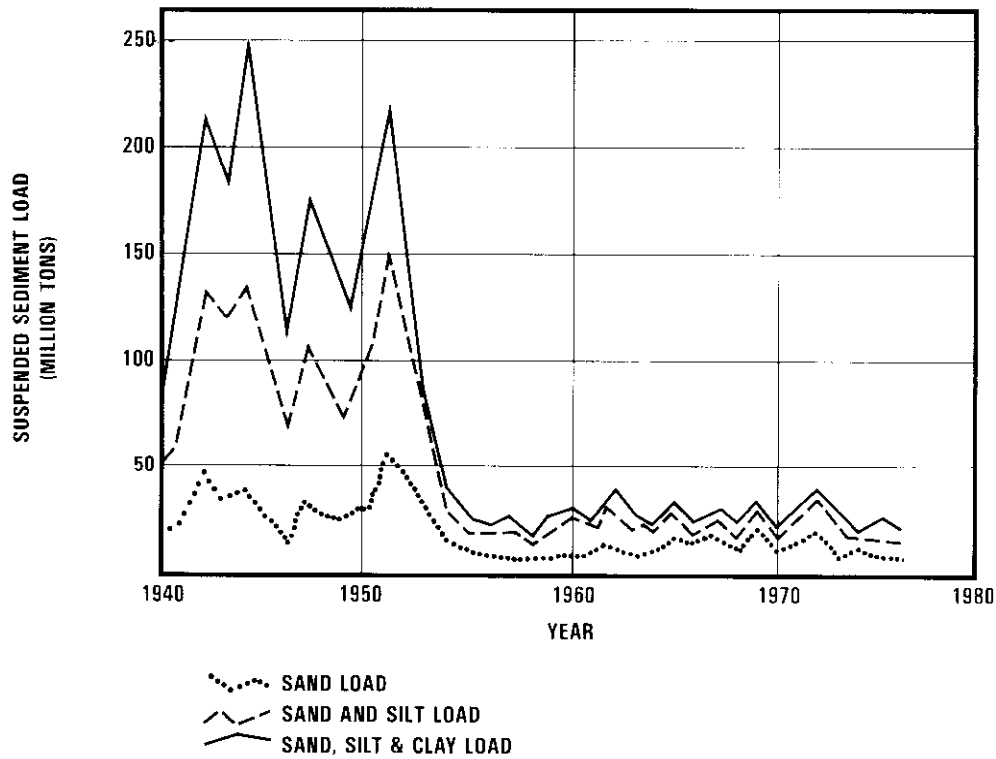


TABLE 2
CAPACITY DEPLETION OF MISSOURI RIVER RESERVOIRS

Project	Long-Term Depletion Estimates (A-F/Yr)		Reservoir Storage (Million A-F)			Measured Depletions		
	Original Estimate	Current Estimate	Original	Latest Date	Survey Volume	Project Age (Yrs)	Annual Rate (A-F/Yr)	Percent Original Storage
Fort Peck	2,197	17,500	19.4	1972	18.9	35	14,400	2.59
Garrison	47,980	38,090	24.3	1979	23.9	26	13,300	1.42
Oahe	39,980	32,290	23.6	1976	23.4	18	15,900	1.21
Big Bend	4,500	4,390	1.90	1983	1.87	20	1,300	1.37
Fort Randall . . .	20,500 ⁽¹⁾	16,600	6.28	1981	5.57	29	24,400	11.26
Gavins Point . .	2,400	2,750	0.54	1979	0.50	24	1,500	6.84

(1) Original estimate included sediment from drainage now controlled by Big Bend

Post-construction resurveys of the sediment ranges at each project have permitted the comparison of observed depletion rates with that estimated from the long-term sediment load estimates, as presented in Table 2. All measured annual depletion rates are less than estimated with the exception of Fort Randall, which in the first six years intercepted the sediment now trapped by Oahe Dam and the first eleven years that by Big Bend Dam. From the post-construction sediment range survey data, it is possible to verify and/or adjust the depletion rates, if warranted. Figure 3 illustrates changes in the depletion rates computed at the time of the given survey. It is apparent that the long-term storage depletion rate is stabilizing and approaching a constant value, although the incremental rates between surveys continue to fluctuate in response to basin runoff.

The total loss of reservoir storage due to sedimentation is an important operating parameter; however, the distribution of these deposits within storage zones in the pool are of particular interest. Storage capacity changes in each pool zone for the six Missouri River projects since closure are shown in Table 3. In interpreting the results, it is well to recognize that the system was not entirely operational until about 1965 when all of the reservoirs finally were filled. A large amount of the material depositing in the lower elevation zones occurred during this prolonged reservoir filling period. The top zone, the exclusive flood control zone, is filled only by rare flood events and is subject to shore line erosion, thus increasing its

FIGURE 3
RESERVOIR DEPLETION RATES

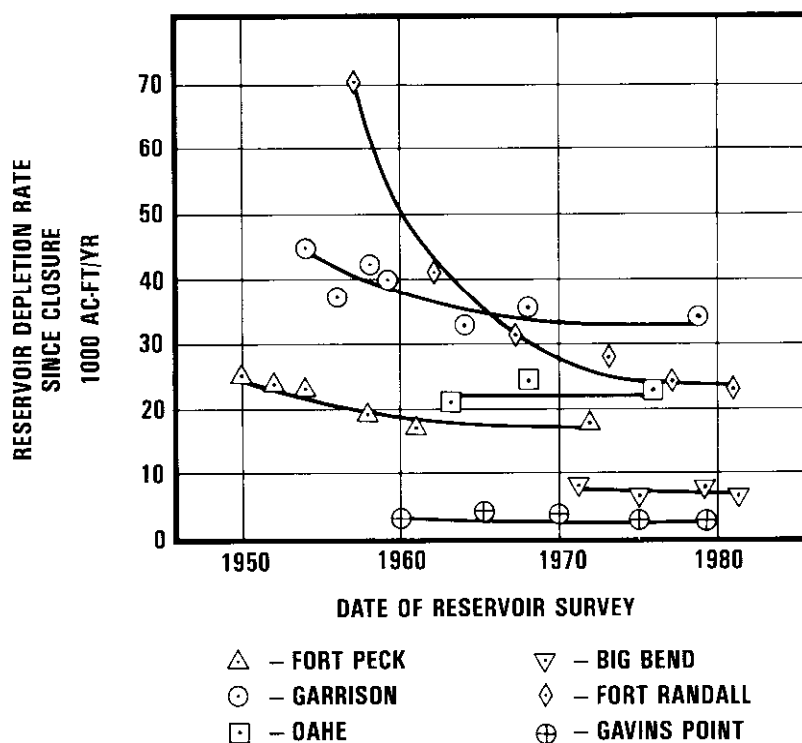


TABLE 3
MISSOURI RIVER RESERVOIRS POOL CAPACITY CHANGES

Project	Pool Zone*	Pool Elev. (Ft m.s.l.)	Original Capacity (Ac-Ft)	Latest Capacity (Ac-Ft)	Change %
Ft. Peck (1937-1972)	1	2250-2246	941,000	977,000	+ 3.83
	2	2246-2234	2,704,000	2,704,000	0.00
	3	2234-2160	11,230,000	10,945,000	- 2.54
	4	2160-	4,537,000	4,283,000	- 5.60
Garrison (1953-1979)	1	1854-1850	1,484,000	1,494,000	+ 0.67
	2	1850-1837.5	4,277,000	4,220,000	- 1.33
	3	1837.5-1775	13,355,000	13,219,000	- 1.02
	4	1775-	5,152,000	4,990,000	- 3.14
Oahe (1958-1976)	1	1620-1617	1,100,000	1,098,000	- 0.18
	2	1917-1607.5	3,179,000	3,186,000	+ 0.22
	3	1607.5-1540	13,808,000	13,603,000	- 1.48
	4	1540-	5,538,000	5,451,000	- 1.57
Big Bend (1963-1983)	1	1423-1422	60,000	60,000	0.00
	2	1422-1420	115,000	117,000	+ 1.74
	4	1420-	1,725,000	1,697,000	- 1.62
Ft. Randall (1952-1981)	1	1375-1365	1,107,000	985,000	-11.02
	2	1365-1350	1,387,000	1,322,000	- 4.69
	3	1350-1320	1,938,000	1,699,000	-12.33
	4	1320-	1,849,000	1,568,000	-15.20
Gavins Point (1955-1979)	1	1210-1208	64,000	61,000	- 4.69
	2	1208-1204.5	102,000	95,000	- 6.86
	4	1204.5-	375,000	348,000	- 7.20

- * 1 Exclusive Flood Control Pool
 2 Flood Control and Multiple Use Pool
 3 Carryover Multiple Use Pool
 4 Inactive Pool

capacity in some of the reservoirs. Wide variations in pool level fluctuations also exist between projects, which have affected the locations of sediment deposits. Gavins Point, for example, normally fluctuates between very narrow limits of 6 ft or less; whereas Fort Randall has fluctuated annually over 18 ft. It is apparent, however, that most deposits exist in a zone that is slightly below the prevailing pool level at the time of the sediment inflow. In the six reservoirs, the majority of the sediment is settled around 10 to 15 miles downstream from the upstream end of the pools, and is concentrated in a 25 to 35 mile reach downstream from this point. Deposition in these zones has caused problems near water supply intakes,

access to the lake and increased flood stages throughout the backwater reach. In the bottom inactive zones of the pool, sediment deposits have not significantly affected the dams and powerhouses at any of the six reservoirs.

DOWNSTREAM TAILWATER TRENDS

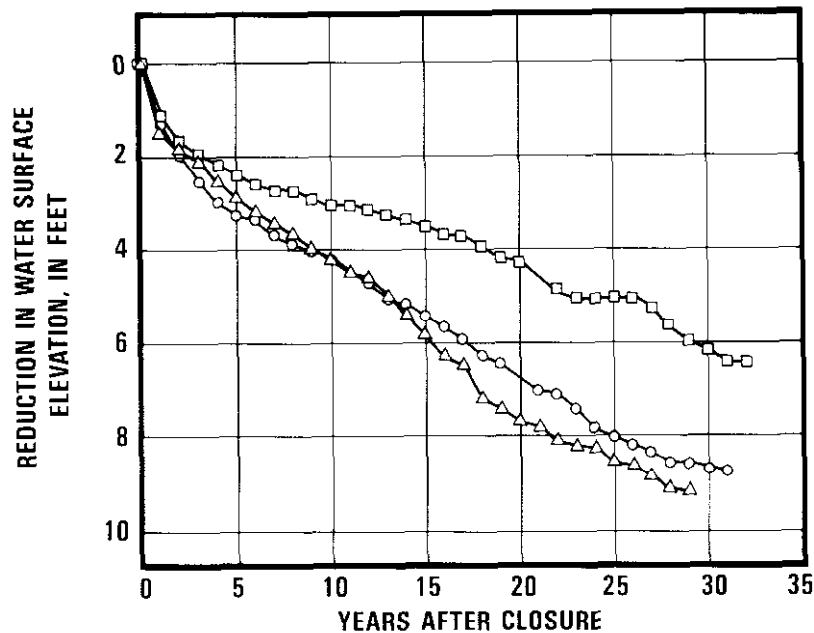
The release of the essentially sediment free water through the dams has resulted in a lowering of the downstream tailwater elevation. Pre-construction estimates predicted that the water surface elevations immediately downstream of the dams would lower a maximum of 15 ft at each project where no fixed downstream control existed. Turbine elevations were set to account for the expected lowering. At Big Bend Dam, the tailwater elevations are controlled by the Fort Randall pool immediately downstream. The tailwater of Oahe Dam which discharges into a short reach of open river will eventually be controlled by Big Bend Reservoir which will limit the eventual tailwater lowering to about six feet. The Fort Peck, Garrison, Fort Randall, and Gavins Point projects, however, discharge directly into open river channel reaches that lie in alluvial deposits up to 100 feet deep, largely composed of fine to coarse sands. The ultimate control to channel degradation below these projects is the presence of small percentages of gravel in the bed material. After several feet of bed lowering, enough gravel will have segregated from the scouring sand and accumulated on the surface of the bed to form an armor layer that eventually will protect the bed from further scour.

Tailwater trends are monitored annually at all projects. The results for Garrison, Fort Randall, and Gavins Point are presented on Figure 4. All of these projects exhibited rapid rates of degradation in the first 2 to 3 years after closure, then gradually leveled off to nearly constant rates. In recent years, however, the rates at Fort Randall increased, possibly attributed to partial loss of armoring material and widening of downstream channel. Field observations of the surface of the streambed below Fort Randall and Gavins Point indicate that coarse materials ranging in size from 2mm to 40 mm are accumulating on the bed surface and are probably responsible for the reduced tailwater lowering trend. A re-acceleration of the degradation rates may be forthcoming; however, with future prolonged releases that may be greater than those already experienced. Tailwater lowering at Garrison Dam continues at an almost uniform rate, with no stabilizing tendency as yet apparent.

BANK EROSION

In the natural river prior to the construction of the dams, there was a balance over the years between erosion of the high banks and the building of new valley lands by sediment deposited during floods, resulting in a continual migration of river channel within the valley. Due to the balance between the erosion and deposition processes, there was no long-term net loss of high valley lands. Since closure of the dams, the operation of the reservoir system has eliminated both the floods and the sediment that were essential for the rebuilding process; however, the process of bank erosion continues. This has resulted in a continuing net loss of high valley lands that are not replaced elsewhere in the valley as in the era before the reservoirs. High valley lands are being converted to river channel and sand bar areas, while the width between high banks continues to widen. This process is transforming

**FIGURE 4
MISSOURI RIVER DAMS
TAILWATER TRENDS**



NOTE: ALL TRENDS ARE FOR A
CONSTANT RELEASE OF
20,000 CFS

○ — GARRISON
□ — FORT RANDALL
△ — GAVINS POINT

the present open river reaches into gradually widening areas of sand bars and channels, occupying increasing proportions of the valley width between the high bluffs.

Aerial photographic surveys taken over the years have been analyzed to obtain estimates of valley lands lost below Garrison, Fort Randall, and Gavins Point Dams. The results of these analyses are summarized in Table 4. The bank erosion below the dams does not occur uniformly along the erodible banklines at the average rates indicated on Table 4. Within each of the open river reaches the destruction of land in any one year is usually concentrated in a few selected locations. These locations of active erosion shift from place to place from year to year. The exact locations where the erosion is likely to occur cannot be predicted in advance; however, over the years every stretch of erodible bankline is potentially subject to active erosion.

The rate of bank erosion also varies with distance downstream from the dam. Table 5 compares observed erosion rates with stage fluctuations from power peaking fluctuations downstream of the projects. Erosion rates experienced to date are generally lower near the dams and have increased with distance

TABLE 4
MISSOURI RIVER BANK EROSION LOSSES

A. GARRISON DAM TO BISMARCK, NORTH DAKOTA			
	Pre-Dam Conditions	Post-Dam Conditions	
	<u>1938-1954</u>	<u>1954-1975</u>	
1. Time Period			
2. Total Erosion Loss, Acres	3561	1691	
3. Erosion Rate, Acres/Yr	222	80	
4. Length of Reach, Miles	79	79	
B. FORT RANDALL DAM TO NIOBRARA, NEBRASKA			
1. Time Period	1946-1953	1954-1975	
2. Total Erosion Loss, Acres	945	1058	
3. Erosion Rate, Acres/Yr	135	47	
4. Length of Reach, Miles	36.3	36.3	
C. GAVINS POINT DAM TO PONCA STATE PARK			
1. Time Period	1930-1956	1956-1979	
2. Total Erosion Loss, Acres	5241	3975	
3. Erosion Rate, Acres/Yr	202	178	
4. Length of Reach, Miles	53.7	53.7	

TABLE 5
EROSION DOWNSTREAM FROM DAMS SINCE CLOSURE

	Distance Below Dam <u>(miles)</u>	Daily Stage Fluctuations due to <u>Power Peaking</u> (ft)	Erosion per Unit Length of River <u>(ac/mi)</u>
Below Gavins Point (1956-1979)	0 to 20	None	42.4
	20 to 40	None	75.4
	40 to 58	None	100.5
Below Fort Randall Dam (1954-1975)	0 to 10	11.0 to 6.5	17.4
	10 to 20	6.5 to 5.6	41.9
	20 to 30	5.6 to 3.3	19.4
	30 to 36	3.3 to 2.6	24.7
Below Garrison Dam (1954-1975)	0 to 20	10.5 to 4.6	8.0
	20 to 40	4.6 to 1.6	22.0
	40 to 60	1.6 to 0.8	20.5
	60 to 79	0.8 to 0.5	18.5

downstream. It is interesting to note that bank erosion below the Missouri River Dams apparently cannot be associated with power peaking since these fluctuations are greatest near the dams and attenuate with distance downstream. In fact, bank erosion losses are greatest downstream from Gavins Point Dam where there is no power peaking and the releases are steady. Downstream bank erosion appears to be associated with picking up the sand load from the bed and the resulting sand bar movements that direct the flow against the banks.

DOWNSTREAM IMPACTS

Completion of the dams with their influence on the downstream river flow and sediment regimes, together with the completion of the open river navigation project downstream from Sioux City, Iowa, has affected channel erosion processes from a considerable distance downstream. As pointed out earlier, the tailwater elevations immediately downstream from the lowest dam, Gavins Point, have lowered over 6 ft since closure. This channel degradation, however, is not restricted to the immediate vicinity of the dam. At 59 miles below Gavins Point, water surface elevations have lowered 2-3 ft, while at Sioux City, Iowa, 78 miles downstream, the lowering is as much as 6 ft. The amount of lowering diminishes with distance downstream from Sioux City, until at Omaha, Nebraska, 195 miles below Gavins Point Dam, no degradation trend is apparent. Upstream from Omaha, the lowering trends continue, but at reducing rates.

CONCLUSION

The six large multi-purpose dams on the main stem of the upper Missouri River have altered the flow and sediment regimes of the river, and completely control the sediment from one of the largest and most sediment productive areas in the continental United States.

The combined impact of sediment entrapment and river flow regulation has had a significant influence not only on the individual projects, but also on the open river reaches downstream. Sediment deltas are forming in the upstream ends of each reservoir, resulting in the loss of reservoir storage capacity and causing higher water surface elevations for flows entering the backwater reaches of the pools. The downstream growth of sediment depositions has not as yet progresses to the dams and powerplants, therefore, no adverse impacts on the hydraulic machinery have been noted. Water released through the dams is influencing the downstream channel and has altered the natural erosion and deposition processed, resulting in changes in the downstream channel characteristics for many miles downstream. Degradation of the downstream channel bed continues, but at a gradually reducing rate. An accurate projection of the ultimate bed elevation of the channel is important to the future management of the system. Bank erosion will continue but at a lesser rate than under the natural condition.

PATTERNS OF SEDIMENT DEPOSITION IN SHALLOW BAYS

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ABSTRACT

The diversion of freshwater and sediment from the Mississippi River to the Atchafalaya River is developing a new land mass (delta lobes) on the south-central Louisiana coast. This paper summarizes an analytical method to predict the patterns of sediment deposition, and the subaerial delta growth in the bay. The method is based on the theory of turbulent plane jets which issue from river mouths and discharge into a quiescent, well-mixed and shallow bay.

The analytical results show that the deposited sediment forms a saddle-shaped bottom. Rapid accumulation of suspended sediments near the river mouth and abrupt decline of sediment deposition away from the mouth are obtained. It is inferred that in a shallow bay the river discharge has a significant influence in sediment deposition patterns over a short distance, and that beyond that distance the river discharge is insignificant.

The analytical predictions are agreeable with predictions made by other approaches. The limitations of the analytical approach are discussed.

INTRODUCTION

The Mississippi River, the largest alluvial river system on the North American continent, has been actively building its delta lobes for at least the past 7,000 years (Coleman, 1976). The river is about 2,300 miles (3,700 km) long, and its drainage area of about 1.29 million miles² (3.34 million km²) encompasses all or parts of 31 states and contains about one-third of the total area of the contiguous states.

The large input of fresh water, with an average annual flow of 542,500 cfs (15,360 cms), and annual sediment discharge of 265 million tons per year (240 billion kg/yr) from the Mississippi River at its delta apex (Coleman, 1976) are undoubtedly the dominant factor in shaping Louisiana coastal land and its associated environment. Most of southeastern Louisiana is made up of overlapping delta lobes, in which the younger deltas overlap older ones that have subsided below sea level (Wright and Coleman, 1974). Seven large scale of delta systems have been identified by various investigators (Kolb and Van Lopik, 1966; Wells et al., 1984). The Atchafalaya River delta is a relatively young and dynamic part of the overall geomorphologic stage of Mississippi River Delta systems (Fig. 1).

The building of the new deltaic landmass in the geographic center of Louisiana has prompted scientific interests not only in predicting the deltaic evolution, but also in assessing the impact on the adjacent coastal embayments.

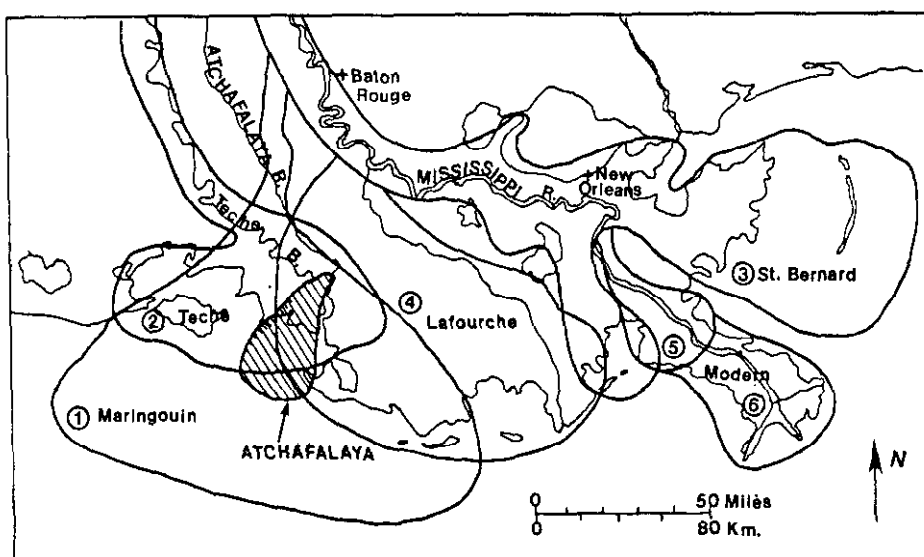


Fig. 1. A Historical Sequence of Delta System in Southeastern Louisiana (After Kolb and Van Lopik, 1966; Wells et al., 1984).

ATCHAFALAYA RIVER-BAY-DELTA INTERACTION

This paper summarizes an analytical approach, derived from the theory of turbulent jets, and used to predict the development of sediment deposition patterns in the Atchafalaya Bay. It is one of several techniques that are being used by the Corps of Engineers to investigate delta growth in Atchafalaya Bay. The Bay is physically contained within an partially enclosed area of 233 miles² (600 km²) as shown in Figure 2. The average depth of the bay is about 5 ft (1.5 m), the water volume is 32.5 million ft³ (0.96 km³), and the average salinity of the bay water is about 0.37 ppt (McAnally and Heltzel, 1978).

The Atchafalaya River is a major distributary of the Mississippi River. Approximately 30% of the combined flow of Mississippi River and Red River flows into the Atchafalaya River Basin (USACOE, 1981). The river flows through an alluvial floodplain characterized by extensive freshwater swamps and numerous small lakes. These swamps and lakes have served as catchbasins trapping much of the sediment before it reached to the bay. The rapid filling of the lakes has accelerated the transport of sediment through the Lower Atchafalaya River and Wax Lake outlets to Atchafalaya Bay (Fig. 2), and, consequently, has resulted in the development of a subaerial delta in the bay. This delta is formed by sediment from the size classes of sand, silt, and clay.

The Atchafalaya delta is building into a shallow, well-mixed enclosed area. The bay-delta system is in an environmental setting of high river discharge, high suspended loads, low tidal range, low wave energy, and low offshore slope. Therefore, our study is focused on fresh water discharging into a shallow bay and the bay's dynamic response to such forcing functions.

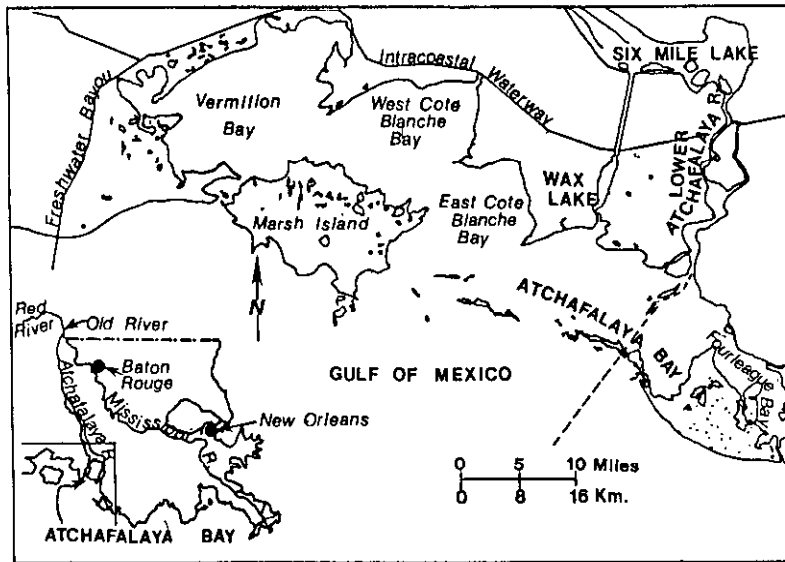


Fig. 2. Atchafalaya Bay and Its Vicinity (After Letter, 1982).

TURBULENT PLANE JET

The river-bay physical system is approximated by a turbulent plane jet issuing into a quiescent bay (Fig. 3). The governing equations of shallow water hydrodynamics are derived from the continuity and momentum equations of turbulent jets (Schlichting, 1968; Abramovich, 1976; Rajaratnam, 1976). These equations are depth- and time-averaged Navier-Stokes equations and are analyzed in detail by Ozsoy (1977), Joshi (1982), and Wang (1984). Under steady state conditions, the bottom-frictional turbulent jet is expressed by

$$\frac{\partial(hu_1)}{\partial x_1} + \frac{\partial(hu_2)}{\partial x_2} = 0 \quad (1)$$

$$\frac{\partial(u_1u_1)}{\partial x_1} + \frac{\partial(u_1u_2)}{\partial x_2} = -\frac{g}{C_z^2} \frac{u_1|u_1|}{h} \quad (2)$$

in which the coordinate system used is defined in Figure 3; where u_1 and u_2 are the velocity components in the longitudinal and lateral direction, respectively; g is the gravitational acceleration; h is the water depth; and C_z is the Chezy's friction coefficient.

SEDIMENT TRANSPORT

Freshwater effluents from river mouths carry sediments in suspension. The diffusion of these materials and their subsequent mixing with the ambient bay water determine their ultimate distribution. The conservation of mass can be described by advection-diffusion processes (Ariathurai et al., 1977). The time-averaged mass transport equation is given by

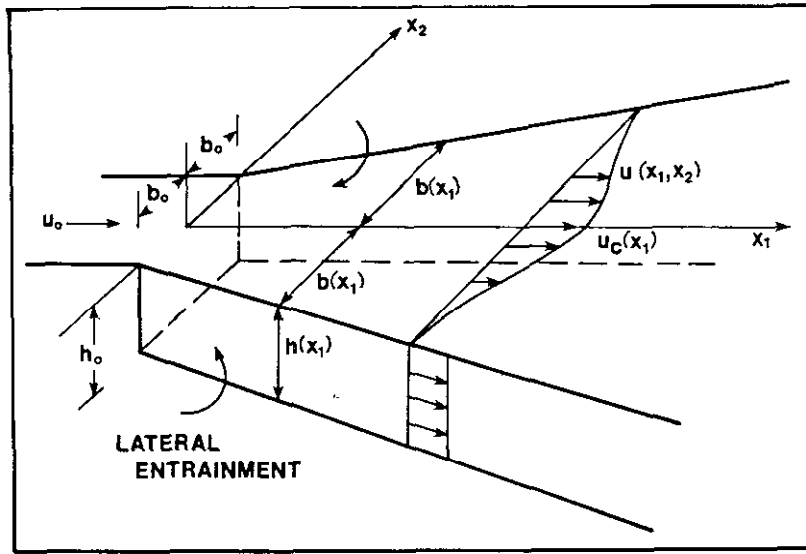


Fig. 3. Representation of River Inflow Discharging into a Quiescent Bay (Wang, 1984).

$$\frac{\partial c}{\partial t} + u_1 \frac{\partial c}{\partial x_1} + u_2 \frac{\partial c}{\partial x_2} = D \left(\frac{\partial^2 c}{\partial x_1^2} + \frac{\partial^2 c}{\partial x_2^2} \right) + S \quad (3)$$

where c is the sediment concentration, D is the diffusion coefficient, and S is the sediment deposition function.

ANALYTICAL SOLUTION

The two important properties of the turbulent jets that simplify the mathematical equations and enable us to derive the analytical solutions for equations (1), (2), and (3), are the self-similarity function (Stolzenbach and Harleman, 1971) and the entrainment function (Allison and Turner, 1959). In this study, the similarity functions for the velocity profile $G(s)$ and the sediment concentration $R(s)$ are expressed as

$$G(s) = (1 - s^2) \exp(-s^2) \quad (4)$$

$$R(s) = (1 - s^2)^{\frac{1}{2}} \exp(-\frac{1}{2}s^2) \quad (5)$$

where s is the normalized lateral distance defined as $x_2/b(x_1)$; and $b(x_1)$ is the half-width of the jet in the longitudinal distance x_1 (Fig. 3).

The velocity v_e caused by the entrainment of the ambient water into the jet is expressed by

$$v_e = e u_c(x_1) \quad (6)$$

where e is the entrainment coefficient; and $u_c(x_1)$ is the jet centerline velocity which is a one-dimensional function of x_1 coordinate.

The profiles of flow velocity and sediment concentration in a horizontal plane are then expressed as

$$u_1(x_1, x_2) = u_c(x_1) G(s) \quad (7)$$

$$c(x_1, x_2) = c_c(x_1) R(s) \quad (8)$$

where $c_c(x_1)$ is the centerline sediment concentration which is also a one-dimensional function of x_1 coordinate.

For convenience and generalization, the variables are normalized as

$$r = \frac{x_1}{b_o}; \quad B = \frac{b}{b_o}; \quad H = \frac{h}{h_o}; \quad U = \frac{u_c}{u_o} \quad (9a)$$

$$C = \frac{c_c}{c_o}; \quad U_{cr} = \frac{u_{cr}}{u_o}; \quad W = \frac{b_o w_o}{h_o u_o}; \quad F = f \frac{b_o}{h_o} \quad (9b)$$

where subscript 0 refers to the values of the variables at the river mouth ($x_1 = 0$); $f = g/C^2$ is the Darcy-Weisbach friction coefficient; w_o is the settling velocity of the sediment particles; and u_{cr} is the critical velocity under which deposition occurs.

Integrating equations (1), (2), and (3) across the width of the jet, using the similarity and entrainment functions, and then simplifying, the normalized forms of continuity, momentum, and sediment dispersion equations are obtained

$$\frac{d}{dr} (HUB) = AHU \quad (10)$$

$$\frac{d}{dr} (HU^2B) = -FU^2B \quad (11)$$

$$\frac{d}{dr} (I_1 HUBC) = -I_2 WBC + I_3 WBC \left(\frac{U}{U_{cr}} \right)^2 \quad (12)$$

subject to the boundary conditions of

$$H = B = U = C = 1 \quad \text{at} \quad r = 0 \quad (13)$$

where $A = 2e/I_0$; I_0 , I_1 , I_2 , and I_3 are the integral constants derived from the similarity functions.

The normalized equations (10), (11), and (12) are types of Bernoulli ordinary differential equations. Their general solutions of the system (linearly varying depth $H = 1 + arb_0/h_0$ with entrainment $A \neq 0$) are obtained by Wang (1985) and expressed as

$$U = H^{-D} \left[\frac{2Ah_0}{ab_0} \frac{1}{2-D} (H^{2-D} - 1) + 1 \right]^{-\frac{1}{2}} \quad (14)$$

$$B = H^{D-1} \left[\frac{2Ah_0}{ab_0} \frac{1}{2-D} (H^{2-D} - 1) + 1 \right] \quad (15)$$

$$C = \exp \left[-\frac{I_2}{I_1} W \int_0^r \frac{1}{HU} dr + \frac{I_3}{I_1} \frac{W}{U_{cr}^2} \int_0^r \frac{U}{H} dr - \ln(HUB) \right] \quad (16)$$

in which $D=f/a$ and a is the slope of the bay bathymetry.

SEDIMENT DEPOSITION PROCESS

Sediment transported by the jet will deposit as the jet velocity drops, and, in the steady flow case, remain on the bed. Those zones where live bed transport occurs are assumed to be in equilibrium. The deposition process is assumed to be describable by a simple rate equation based on work by Krone (1962) with clays. In this approach, deposition occurs only when the flow speed drops below a critical value. The deposition rate is assumed to be proportional to the concentration, fall velocity, and ratio of flow speed to critical speed. The sediment thickness, $d(r,s)$, is assumed to be varied linearly with time (Engelund, 1976) and is given by

$$d(r,s) = \frac{1}{\rho_s} \int_0^t w_o c(r,s) \left[1 - \frac{u(r,s)^2}{u_{cr}^2} \right] dt \quad (17)$$

where ρ_s is the bulk density of the sediment, and t is the time from the beginning of the deposition process, and

$$u(r, s) = u_o (1 - s^2) \exp(-s^2) U(r) \quad (18)$$

$$c(r, s) = c_o (1 - s^2)^{\frac{1}{2}} \exp(-\frac{1}{2}s^2) C(r) \quad (19)$$

PATTERNS OF SEDIMENT DEPOSITION

The special cases for the general solution of equations (10), (11), and (12) are derived in detail by Wang (1985). The sediment concentration for various cases are summarized below:

Case 1: Constant depth without entrainment ($H = 1, A = 0$)

$$C(r) = \exp \left\{ \frac{I_2}{I_1} \frac{W}{F} [1 - \exp(-Fr)] + \frac{I_3}{I_1} \frac{W}{F U_{cr}^2} [1 - \exp(-Fr)] \right\} \quad (20)$$

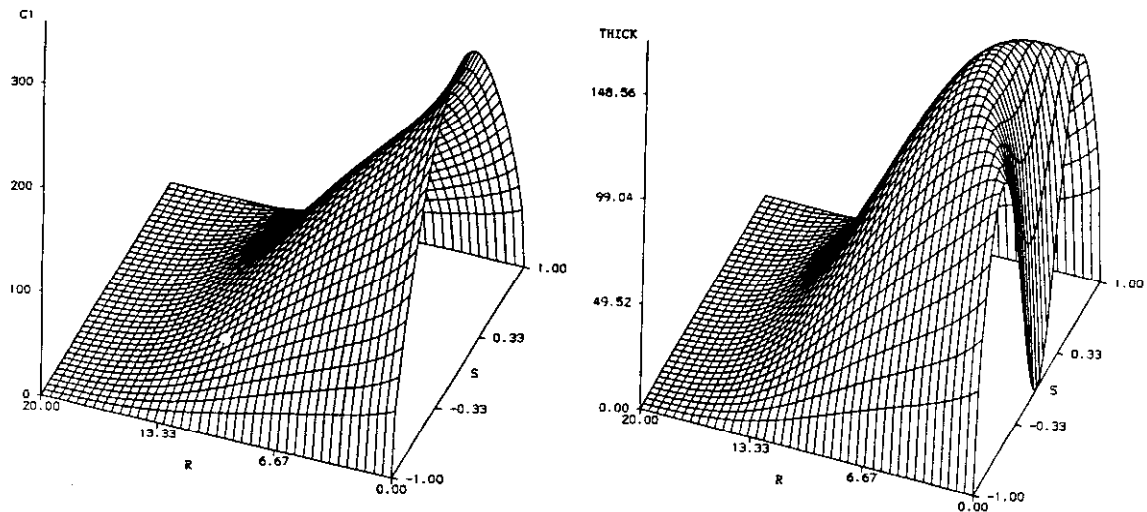
Case 2: Constant depth with entrainment ($H = 1, A \neq 0$)

$$C(r) = \exp \left[-\frac{I_2}{I_1} \frac{W}{F} \int_0^r \frac{1}{U} dr + \frac{I_3}{I_1} \frac{W}{U_{cr}^2} \int_0^r U dr - \ln(UB) \right] \quad (21)$$

Case 3: Linearly varying depth without entrainment ($H \neq 1, A = 0$)

$$C(r) = \exp \left[-\frac{I_2}{I_1} \frac{W}{F} (1-H^D) + \frac{I_3}{I_1} \frac{W}{F} (1-H^D) \right] \quad (22)$$

Figures 4a and 4b display the three-dimensional plot of sediment concentration and sediment thickness for Case 4 (linearly varying depth with entrainment $H \neq 1, A \neq 0$).



(a) Sediment Concentration in ppm. (b) Sediment Thickness in cm.
Fig. 4. Case 4: Linearly Varying Depth with Entrainment ($H \neq 1, A \neq 0$).

PREDICTIONS OF DELTA GROWTH

The physical dimensions of the Atchafalaya River-Bay system are approximated by simple geometry of the domain for analytical solutions (Wang, 1985). The cross section of the Lower Atchafalaya River Outlet is taken as rectangular with a width $2b_0 = 1,000$ m and a depth $h_0 = 2$ m. The bathymetry slope of Atchafalaya Bay is $a = 0.0001$. The river discharge has a flow velocity $u_0 = 1$ m/sec and a sediment concentration $c_0 = 300$ ppm. The settling velocity of the sediment particles is $w_0 = 0.05$ mm/sec. The sediment has a bulk density of $\rho_s = 400$ kg/m³. The bottom friction coefficient is assumed to be $f = 0.001$, and the entrainment coefficient $e = 0.0075$ is used (Engelund, 1976).

A numerical procedure has been developed to estimate the areal and volume extent of delta growth in both space and time. Guided by the dimensional plot of sediment deposition patterns (Fig. 4), the space steps are selected at the beginning of the computation. The time steps are not known a priori, but depend upon the analytical solutions themselves as given by equations (14), (15), (16), and (17). Detail computational procedures are given by Wang (1985). Figure 5 shows our predictions of subaerial land in Atchafalaya Bay compared with the predictions made by other studies (Shlemon, 1972; Letter, 1982; Wells et al., 1984). A contour configuration of subaerial delta in the bay is displayed in Figure 6 for the general solution of the system (Case 4: linearly varying depth with entrainment $H \neq 1$, $A \neq 0$).

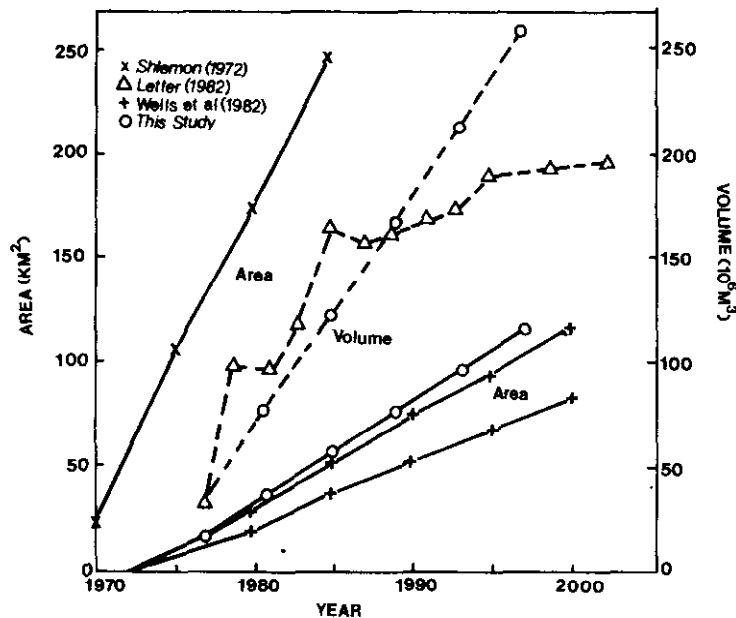


Fig. 5. Prediction of Subaerial Land in Atchafalaya Bay, Louisiana.

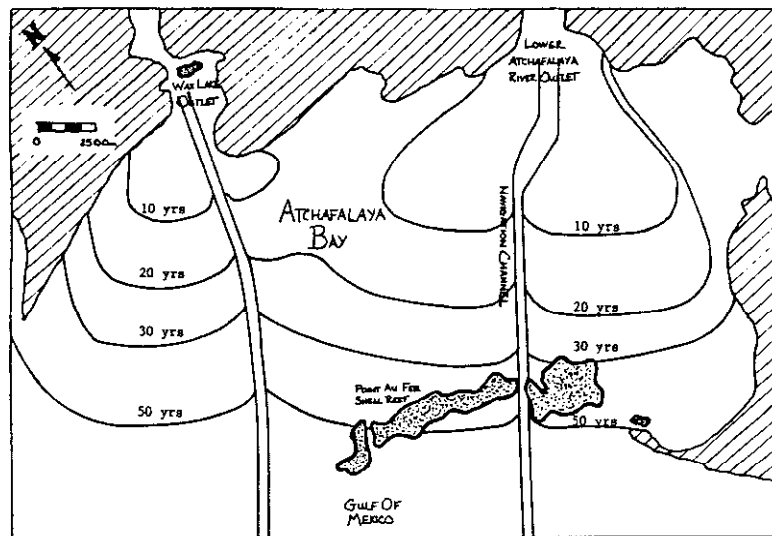


Fig. 6. Contour Configuration of Subaerial Delta in Atchafalaya Bay, Louisiana.

SUMMARY

The river-bay-delta evolution is a complex phenomenon. The analytical results summarize in this paper are simplified representation of a real system, in an attempt to identify the important features of physical processes. Although our analytical predictions are close to predictions by others, the analytical method has its limitations. The governing equations are applied to shallow-water environment without waves and tides. The results are obtained by the assumptions used in the simplification and initial and boundary conditions imposed in the problem solutions.

However, the readily obtainable analytical solutions make the approach rather attractive in comparison with more complicated computer models. The analytical approach serves as a tool to predict the gross estimate of a newly developed subaerial delta growth in a well-mixed and shallow bay.

Our analytical results reveals that the most striking phenomenon of river-delta formation in the shallow bay is that the river discharge has a significant influence on the sediment deposition patterns within a short longitudinal distance, and beyond that distance the river discharge is relatively insignificant.

ACKNOWLEDGMENT

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COLORADO RIVER SEDIMENT TRANSPORT IN GRAND CANYON

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ABSTRACT

The generating units at the Glen Canyon Dam Powerplant will be uprated and rewound by 1987. This means that the powerplant will have an increased capacity to fluctuate flows and thus generate more revenue. Because of this new capacity, studies were initiated to evaluate the short and long term environmental impacts to recreation, fisheries, vegetation, and beach erosion of the Colorado River through Grand Canyon National Park.

The U.S. Geological Survey, under an agreement with the U.S. Bureau of Reclamation, collected suspended sediment samples, bed material samples, velocity measurements, and cross section data. These data were collected at five sampling stations on the Colorado River and three tributary stations between Lees Ferry and Diamond Creek. The sampling period contains two flow conditions: one of declining large discharges ranging from 90,000 to 21,000 ft³/s, from June to December 1983, and the other will consist of a daily peaking operation at Glen Canyon Dam with flow ranging from 3,000 to 28,000 ft³/s from October 1985 through January 1986.

The paper compares sediment transport versus discharge rating curves produced by the following six transport equations: Modified Einstein, Velocity-Xi Adjusted Einstein (Bureau of Reclamation), Meyer-Peter and Muller, Ackers and White, Toffaleti, and Yang.

INTRODUCTION

Sediment sampling on the Colorado River Between Glen Canyon Dam and Lake Mead is an important aspect of the Glen Canyon Environmental Study. The Glen Canyon Environmental Study was funded and coordinated by the USBR (U.S. Bureau of Reclamation) in cooperation with the National Park Service, U.S. Fish and Wildlife, U.S. Geological Survey, and Arizona Game and Fish Department. A more complete description of the many facets of the sediment studies is given in a companion paper included in these proceedings on "Sediment Studies in Grand Canyon Arizona" by J. B. Graf, E. L. Pemberton, and J. C. Schmidt.

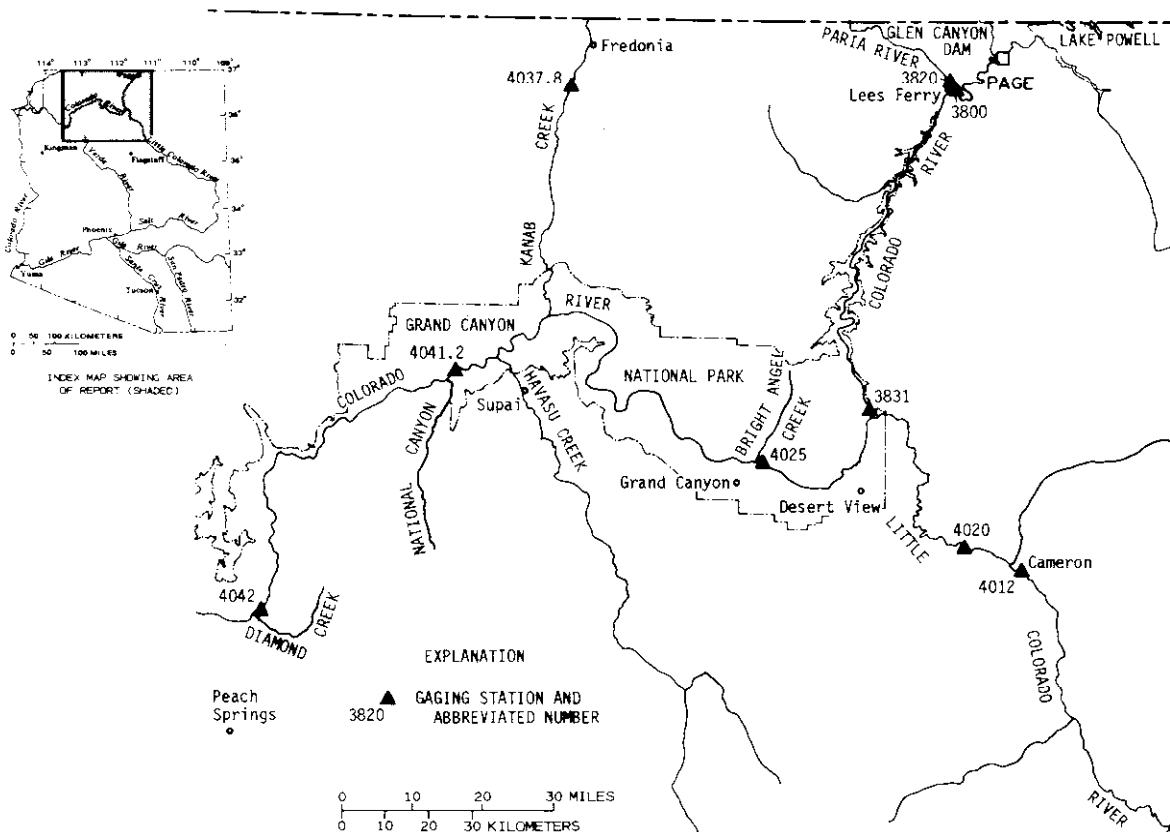
This paper discusses the sediment data collection and analysis program from June to December 1983 and from October 1985 to January 1986. The objective of the sediment sampling program is to collect sufficient channel hydraulic data and sediment samples to establish sediment rating curves at strategic points in 225.5 miles (363 km) of the Colorado River from Lees Ferry to Diamond Creek. The sediment rating curves at each sampling site would serve two purposes: (1) control points in analyzing changes in sediment transport under alternative flow-release patterns below Glen Canyon Dam and

(2) calibration of the Sediment Transport and River Simulation (STARS) model which includes determining the best predictive equation to use for comparing changes in sediment transport under alternative release patterns.

The five Colorado River sampling sites divided the river below Lees Ferry into four reaches and made use of the two existing gages "at Lees Ferry" and "near Grand Canyon" (see figure 1). One sampling station was located just above the mouth of the Little Colorado River which is the major tributary downstream from the Paria River. The sampling gage "above Diamond Creek" represented an outflow station from the Grand Canyon and the gage "above National Canyon," provided an intermediate station. An important part of the program was collection of sediment samples on the three major tributaries, Paria River, Little Colorado river, and Kanab Creek. These three tributaries were sampled in previous years and the new samples will provide a check with the earlier data.

FIELD DATA

Collection of channel hydraulic and sediment data at each sampling site was unique due to remoteness of the area and the record high flows since the closure of the dam, in 1963. The flows consisted of a falling stage hydrograph ranging from about 91,000 ft³/s on June 30, 1983, "at Lees Ferry" to about 21,000 ft³/s on December 14, 1983, "near Grand Canyon".



The location of the sediment sampling station and reach of river involved are listed below:

<u>Gage</u>	<u>River Mile</u>	<u>Reach Length</u>
Colorado River		
At Lees Ferry	0.0	
Above Little Colorado River	61.0	61.0 mi (98 km)
Near Grand Canyon (Phantom Ranch)	87.5	26.5 mi (43 km)
Above National Canyon	166.0	78.5 mi (126 km)
Above Diamond Creek	225.5	59.5 mi (96 km)
Paria River near mouth	0.8	
Little Colorado River near Cameron	61.3	
Kanab Creek near Fredonia	143.4	

The 1983 sampling was conducted by teams of two people spending ten days at each gage before moving to a new location. The goal for the data collection schedule was as follows:

- First day of every 10-day crew rotation
 - Suspended sediment - 20 to 25 verticals (EWI or Equal-Width-Increment Method)
 - Bed Material - 20 to 25 points
- Remaining 9 days
 - Alternate from day to day
 - Day 1 Discharge Measurement
 - suspended sample (10 to 15 verticals-EWI Method)
 - Day 2 Two sets of suspended samples
 - (10 to 15 verticals-EWI Method)

The major change in the sampling procedure for the unsteady flow period (planned for 4 months beginning in October 1985) will be to measure the discharge as close in time as possible to the suspended sediment sample. Due to consistency in bed material gradations in 1983, a lesser number of bed material samples will be collected in the 1985-86 sampling.

SEDIMENT TRANSPORT

The linking of several different types of data was necessary to compute the sediment transport at each location. The U.S. Geological Survey collected suspended sediment samples, bed material samples, velocity measurements, and river cross section data at each gage. The major complication was in linking all of the data needed to make one sediment transport computation for a given location and time. The complication arose from the different frequency and time of data collection. Since suspended sediment concentrations were thought to vary the most rapidly with time, samples were collected about three times every 2 days. The discharge measurements and bed material samples were taken about once every 2 days. This means that there are more measured suspended sediment concentrations than discharge measurements or bed material samples. The process of linking data and computation of sediment transport is illustrated by the flow chart shown in figure 2.

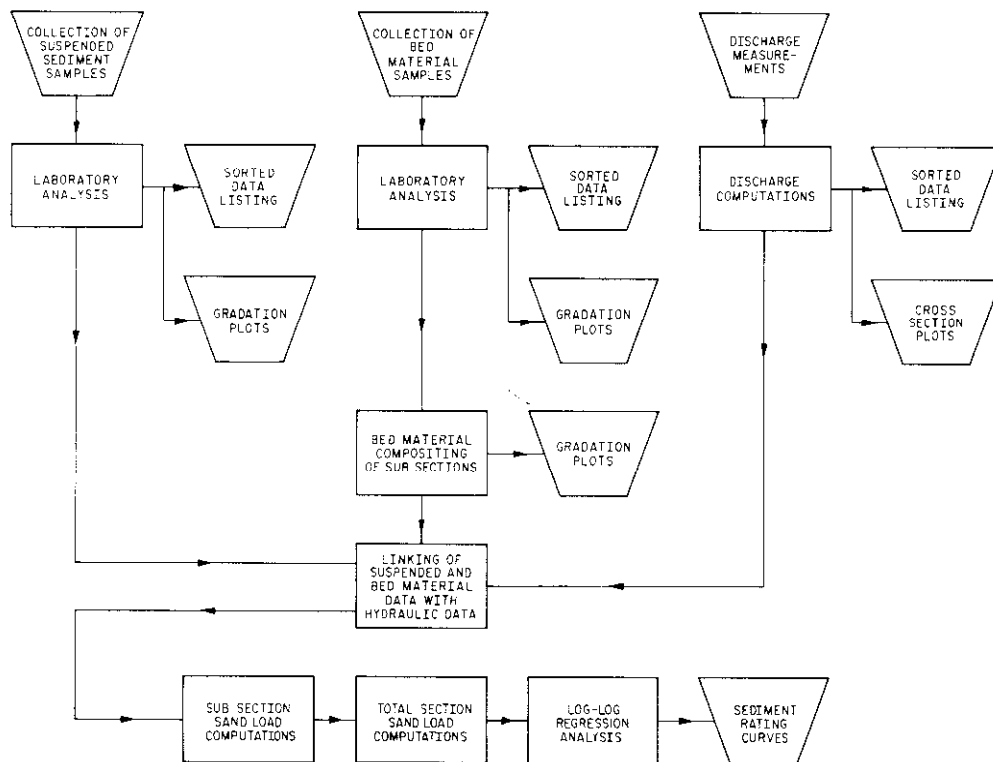


FIGURE 2—DEVELOPMENT OF SEDIMENT RATING CURVES

A computerized data management system was developed to combine the tremendous variety and volume of data for computation of sand load at each of the five gaging stations and for different transport equations. The data management system made it possible to sort and plot the data in a variety of ways. It also eliminated tedious hand computations and the need for reentry of data for use with the various programs. Without the aid of the data management system the amount of work accomplished would have been both time and cost prohibitive.

Computation of sand transport by the Modified Einstein Method (Colby and Hembree, 1955) was adopted to be the total sand in transport at the time of the suspended sample. With this Modified Einstein transport as a control, several of the many different predictive sediment transport equations were applied to the Colorado River data. Total sediment transport was computed using each of the following five predictive transport equations:

- Velocity-Xi Adjusted Einstein, Bureau of Reclamation, (Einstein, 1950; Pemberton, 1972)
- Meyer-Peter and Muller (Meyer-Peter and Muller, 1948)
- Ackers and White (Ackers and White, 1973)
- Toffaletti (Toffaletti, 1969)
- Yang (Yang and Molinas, 1982)

Bed Material Composites

Plotting the bed material size gradation for each sample revealed some drastic changes in bed material sizes ranging from sand to gravels and cobbles across the cross section at the five sampling locations. However, there appeared to be only minor changes with time during the sampling period at a given point. The bed material samples were composited by averaging data representing a predetermined portion of the cross section and time period. Each bed material composite was applicable to a particular river gage, cross section limits, and time period with up to four bed material composites in a cross section. A bed material composite could apply to two separated portions of the cross section. An example of compositing bed material into four different size gradations is shown in figure 3.

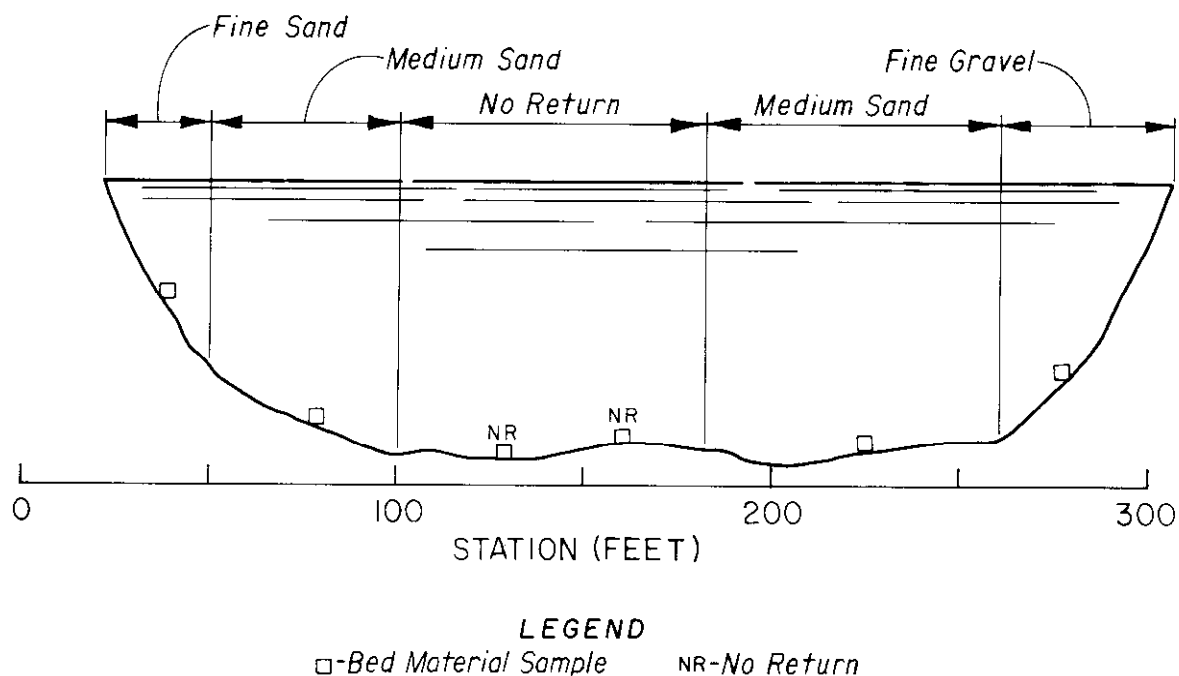


FIGURE 3—TYPICAL CROSS SECTION AND BED MATERIAL COMPOSITES

One problem with measuring bed material on the Colorado River is that the BM-54 sampler, when contacting gravel or cobble-size bed material, closes on contact with the bed and no material is collected. When this occurred, it was identified as a "no return" or "NR" sample. In some cases, samples were collected near the interface where the bed material was changing from sand to gravel material and these samples had a few particles of coarse gravel and a small amount of sand. Most of the bimodal type of samples were not used in computing a composite size gradation.

The linking of the composite bed material size gradation with the suspended sediment data was always done by date. Because the bed material varied across the cross section as shown in the sketch, the sand load was

computed separately between the station limits of each bed material composite for a given sampling station, date and time. No composites were made where the bed material was identified as "no return". The total sand transport for this portion of the channel was assumed to be the measured suspended sand load.

Hydraulic Properties

The hydraulic properties used to compute sediment transport were obtained from the discharge measurement collected closest in time to the suspended sediment sample, with the closest gage height, and closest cross section shape.

In linking the channel hydraulic data, it was assumed that the velocities from the discharge measurement are the same as for the suspended sediment sample. If the gage height for discharge measurement and suspended sample were different, the discharge was recomputed using the gage height of the suspended sediment sample and the velocities from the discharge measurement. For cases where the gage height of the suspended sediment sample was greater than the gage height of the discharge measurement, the cross section boundary was extended based upon data from the discharge measurement with the highest gage height.

The bed material changes within the cross section (figure 3) provided the basis for subdividing the cross section in the sediment transport computations. The discharge, area, wetted perimeter and top width were computed for the cross section portion between the station limits of each bed material composite.

Suspended Samples

The suspended sediment concentration and size gradation represented the results of an EWI (Equal-Width-Increment) Method of sampling. This permitted the compositing of all verticals sampled into one sample for laboratory analysis. Therefore, suspended sample results are reflective of transport within the entire cross section and are assumed to be uniform across the cross section for a given time. This assumption is considered good because of the interaction with the cross section caused by irregular bed forms, channel velocities, and turbulence.

Modified Einstein Method

The first step was computation of total sediment transport by the Modified Einstein Method. In these computations, as well as the other five predictive equations, the hydraulic properties were determined for a subdivided portion of the cross section as defined by the bed material composites. The method used was the Bureau of Reclamation's version of Modified Einstein whereby "Z" values (defined as the theoretical exponent for vertical distribution of sediment) are computed for two or more sand sizes with sufficient material found both in suspension and in the bed. After computing the "Z" values for two or more reference sizes, the "Z" values for all other sizes are computed from a least squares regression of "Z" versus fall velocity. The "Z" values used as reference size in the correlation are used rather than the computed value of "Z" from the regression equation.

The initial assumption in the "Z" computations was made by computing "Z" values for all size fractions having 1 percent or greater material in both suspension and bed material. After the first run, this percentage was changed in some cases in order to complete a computation and/or to have a better correlation for extrapolation purposes.

An example of a total sediment transport computation for a subdivided cross section by the Modified Einstein Method for the "above Little Colorado River" gage on October 10, 1983, is shown in table 1.

Table 1
COLORADO RIVER ABOVE LITTLE COLORADO RIVER
MODIFIED EINSTEIN COMPUTATION

DATE	TIME	GAGE HEIGHT	CONCENTRA- TION	BED MATERIAL DATES		HYDR DATE	HYDR SLOPE	N VALUE
				FROM	TO			
10/10/83	845	2685.28	167.	7/12/83	12/13/83	10/10/83	.0002800	.035

	Composite 1	Composite 2	No Return	Total
Stations (ft)	50 to 130 330 to 380	380 to 450	130 to 330	62 to 409
Discharge (ft ³ /s)	5925.	228.	20714.	26868.
Sampled (tons/day)	2672.	103.	9340.	12115.
Sand Load (tons/day)	3422.	107.	8443.	11973.
Total Load (tons/day)	3679.	117.	9340.	13136.
Area (ft ²)	1919.	187.	3760.	5866.
Velocity (ft/s)	3.09	1.22	5.51	4.58
Top Width (ft)	118.0	29.0	200.0	347.1
Wetted per (ft)	126.3	31.3	200.3	357.9
Hyd Radius (ft)	15.20	5.97	18.77	16.39
Hyd Depth (ft)	16.26	6.44	18.80	16.90
D35 (mm)	.28	.16		
D50 (mm)	.31	.20		
D65 (mm)	.35	.24		
D90 (mm)	.45	.39		

Predictive Equations

Upon completion of the Modified Einstein computations, the same bed material size analyses and hydraulic parameters were used to compute the transport by five different predictive equations. For each computation by a predictive equation the results were presented in tables similar to table 1, with the composites added together for total sand transport.

SAND LOAD RATING CURVES

The 1983 data were used to develop sand load rating curves for the six sediment transport computations at each of the five gages. Data from the gage on the Colorado River above Little Colorado River was selected to illustrate the scatter in the sand loads computed by Modified Einstein Method (figure 4). The same scatter in Modified Einstein data points occurred at all five gages (figure 5) and was traceable back to extremes in the suspended sediment sample results. Extremes in suspended sediment

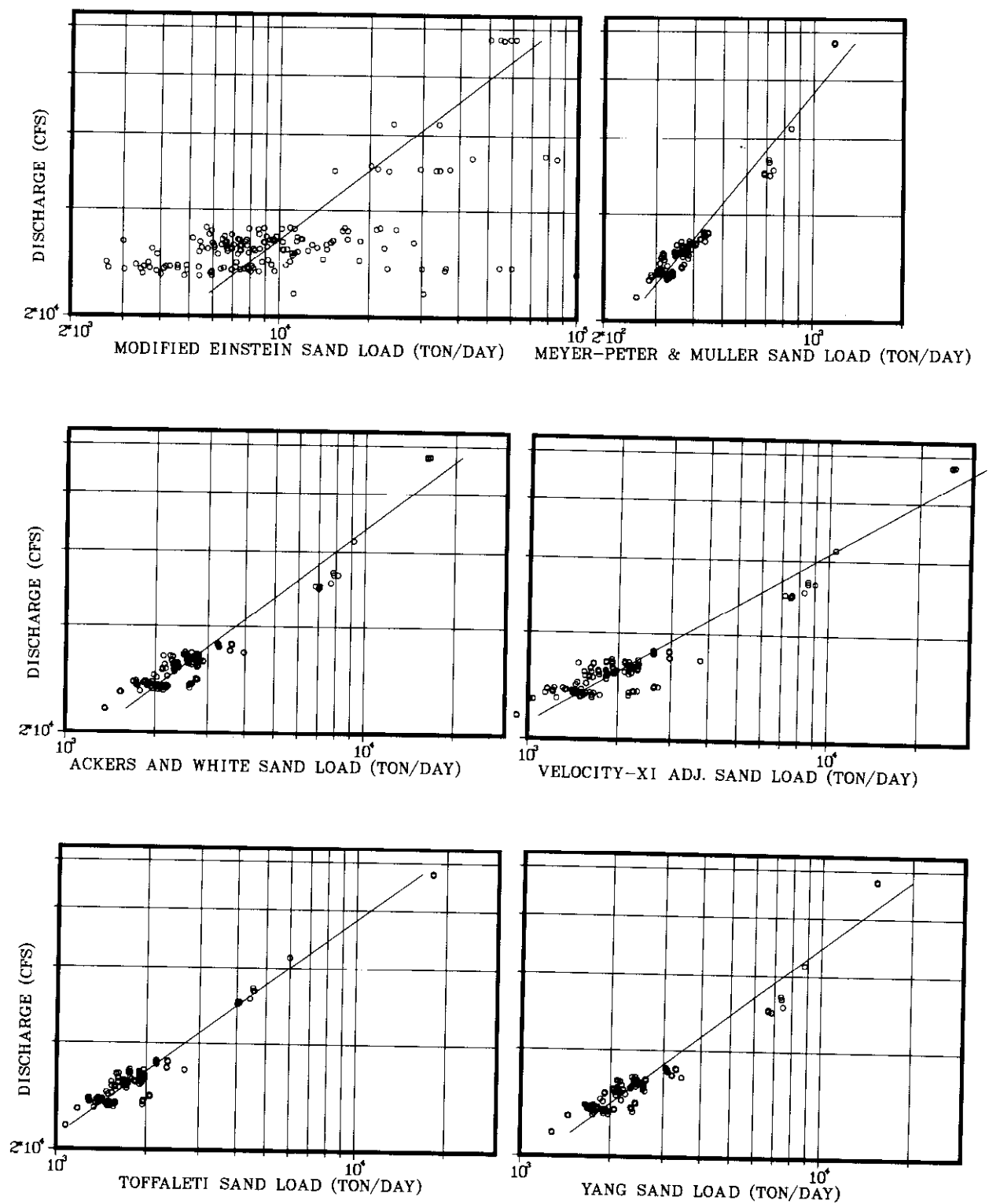


Figure 4 - Sand Load Rating Curves, Colorado River
Above Little Colorado River

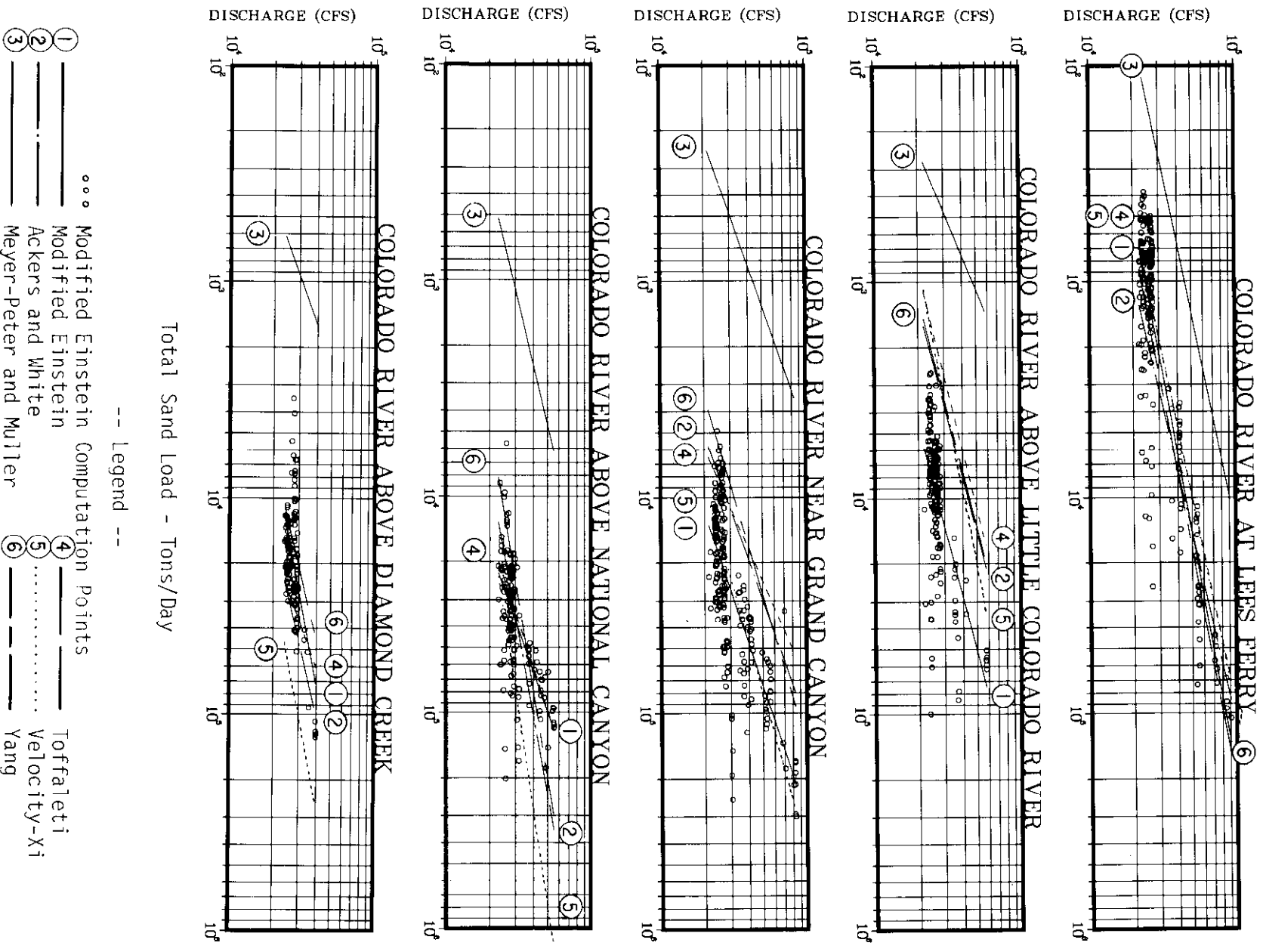


Figure 5 - Colorado River Sand Transport Rating Curves

samples on the Colorado River are not unusual because of tributary or side drainage inflow of sediment in addition to filaments of sediment movement in suspension caused by sand dune or sand bars influencing instantaneous samples. Sand load rating curves from the predictive equations which rely solely on bed material size gradation and channel hydraulic properties, as anticipated, had much less scatter with a better defined regression relationship (figure 4).

The regression curves are shown in figure 5 for all six transport equations at each of the five gages. Because of the limited range of water discharge in 1983 no definite conclusion could be reached with regards to the most applicable equation. The one exception was the Meyer-Peter and Muller equation which consistently computed low sand loads. The best predictive equation can be determined after analysis of the 1985-86 data. This sampling period will have fluctuating discharges in a range consistent with a normal powerplant release pattern.

CONCLUSIONS

The development of a computerized data management system, that linked the large amount of sediment and channel hydraulic properties together for computing sediment transport at five sampling gages on the Colorado River, provided an efficient and reliable method to complete the computations and present the results.

The procedure for compositing the data into subdivided sections for total sediment transport at a cross section permitted sediment transport computations under some extreme variations in basic data. The tributary inflow and temporal changes in suspended sediment loads contributed to the scatter of data points at all of the five sampling gages. In some cases, the changes in suspended sediment loads appeared to be influenced by variations in bed material from sands to gravel and up to cobble-size sediments. The predictive equations, even with the scatter in data points, provide good mean sediment loads for the period of sampling. The limited range in discharges basically greater than 20,000 ft³/s in 1983 emphasized the need to collect sediment data on the normal fluctuating powerplant releases for the 4-month period beginning in October 1985. Discharges for the 1985-86 sampling period ranging from 3,000 ft³/s up to about 31,500 ft³/s will provide sufficient data on sand transport to define a rating curve under variable power plant releases at Glen Canyon Dam. Sand transport at the five gaging stations is being used to calibrate the Bureau of Reclamation's sediment routing model (STARS) for predicting long term sediment transport and channel changes under variable release patterns. These studies on sediment transport will be used in conjunction with other studies on sandbar changes and volumes of sand available in the streambed for projecting the short and long term environmental impacts of sediment movement through Grand Canyon National Park.

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PROBABILITY DISTRIBUTIONS FOR BEDLOAD TRANSPORT

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ABSTRACT

Bedload transport rates at a fixed point fluctuate widely with time, even with a steady-flow condition. When the bed is formed into dunes, the transport rate varies from zero or near zero in the troughs to a maximum at the crests. Although documentation is sparse, rate fluctuations at a point also seem to occur even when bed forms are not distinct.

An existing probability distribution function can be used to describe bedload transport for steady flow over a dune bed. According to this distribution, transport rates vary from zero to four times the mean, and 60 percent of all rates are less than the mean. Several sets of measured transport rates from dune beds show good agreement with this distribution even though these sets were collected from a variety of bed materials using different types of samplers.

The maximum transport rate of four times the mean and the resulting shape of the distribution are specific to the geometric assumptions made in the derivation. The derivation can be extended, by using general rather than specific geometric assumptions, to yield expressions for a family of distribution functions. This generalization results in two simple equations for determining the maximum relative transport rate and the variance of any assumed distribution. The variance of a given distribution can then be used to examine the accuracy of sample sets consisting of various numbers of observations. Since these functions are free of specific geometric constraints, their potential applications are not restricted to specific bed forms.

INTRODUCTION

Both the spatial and temporal scales of the inherent variability in bedload transport are orders of magnitude greater than for the variability associated with commonly measured hydrologic variables such as suspended-sediment concentration and stream velocity. With portable accumulating-type bedload samplers, such as the Helley-Smith (1971) sampler, sampling times are approximately equivalent to the period of random bedload fluctuations and are considerably less than the period of oscillations caused primarily by bed-form migration. Thus, bedload samplers do not integrate over time long enough to average out random fluctuations in a manner analogous to current meters or integrating-type suspended-sediment samplers. In order to account for the extreme inherent temporal variability in bedload transport rates, sampling procedures must be based on knowledge of the characteristics and duration of the expected variability.

Inherent variability for the case of a dune bed with nearly steady flow is described by an existing probability density function and corresponding distribution function (Hamamori, 1962). These functions have been shown to

accurately describe measured results from field and laboratory studies with a variety of bed materials and different bedload samplers (Carey, 1985; Hubbell et al., 1985). In general, these field and laboratory studies were conducted during nearly steady-flow conditions and with distinct bed forms. Thus, sufficient information exists to determine the accuracy of sampling procedures for cases where the existing functions apply. This paper presents a generalized form of the existing probability functions and illustrates how distribution characteristics, and inferred sampling procedures, change for specific applications of the generalized functions. Mathematical derivations will not be presented.

TEMPORAL VARIABILITY

Observations of large fluctuations in transport rates measured at a fixed point over time have been reported in the literature since at least the 1930's. Ehrenberger (1931) reported large fluctuations in data obtained from the Danube River, and Einstein (1937) was able to identify several distinct oscillations with different frequencies in data obtained from a laboratory flume. These large temporal fluctuations in measured bedload transport rates have been reported for a range of bed materials and for a variety of samplers (table 1).

Table 1. Selected references that present data on temporal variability of bedload transport rates measured at a fixed point.

Author	Date	Material	Sampler	Bedforms
Ehrenberger	1931	Gravel	Basket-type (field)	Not reported
Nesper	1937	Gravel	Basket-type (field)	"Waves"
Einstein	1937	Gravel	Basket-type (lab)	Not reported
Hamamori	1962	Sand	Flume (trap)	Dunes
deVries	1973	Sand	Arnhem (field)	Dunes
Emmett	1975	Gravel	Helley-Smith (field)	Not distinct
Hubbell, et al.	1985	Gravel	Helley-Smith (lab)	Dunes
Hubbell, et al.	1985	Gravel	Flume (trap)	Dunes
Carey	1985	Sand	Helley-Smith (field)	Dunes

Typical temporal records from both field and laboratory studies are illustrated in figure 1. Transport rates from laboratory traps are usually represented by continuous vertical bar charts because sampling is nearly continuous. In the field, several minutes may elapse between sequential samples as the sampler is retrieved, emptied, and lowered back to the stream bed. Therefore, sampled rates are usually connected by straight lines which serve more as a visual aid than a confident interpretation of bedload transport between samples.

The oscillatory nature of bedload transport is easily discernible in the laboratory data. Carey (1985) has shown that simple smoothing techniques can be applied to sufficiently large sets of field data to make rate oscillations more apparent. Most field data sets, however, consist of statistically few observations, which limits their use in statistical analysis. For the case of a dune bed, transport rate oscillations correlate with the passage of bed forms beneath the sampling point (Hamamori, 1962; Hubbell et al., 1985). By observing the behavior of dunes in a sand-bed flume, Hamamori (1962) derived

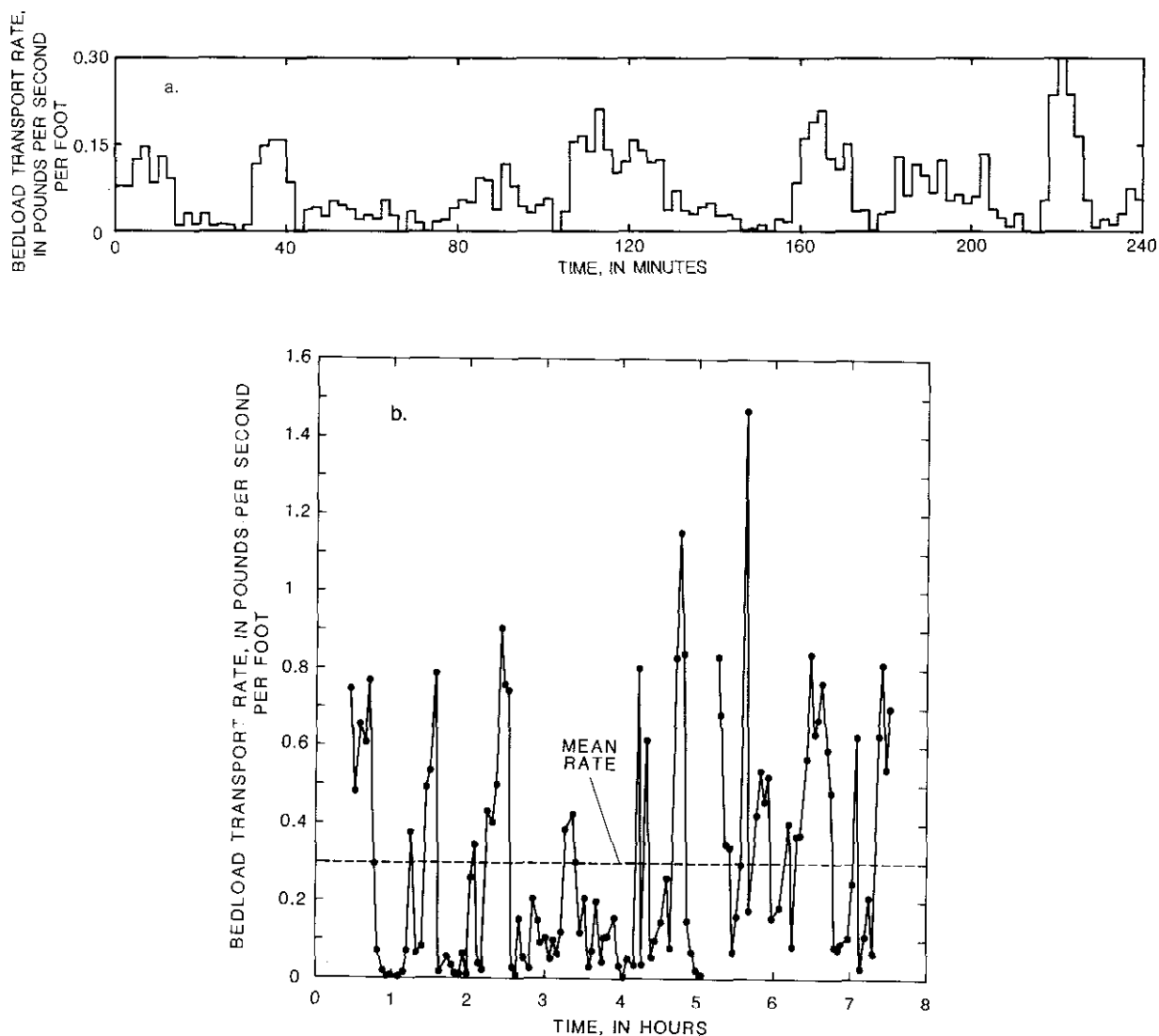


Fig. 1. Examples of temporal variability in bedload transport rates from (a) laboratory and (b) field.

a probability density function of relative transport rates for a dune bed during steady flow. A relative transport rate is simply an individual rate divided by the mean rate.

PROBABILITY FUNCTIONS

Hamamori (1962) first looked at a simple triangular bed form and noted that if the bed form moves down the channel without changing its shape, then the transport rate must increase linearly from zero at the upstream trough to a maximum at the crest. Because the transport rate increases linearly, the mean rate is one half the maximum rate, or in terms of relative rates, (y), the ratio of the maximum rate to the mean rate, (Y), is two (fig. 2a). The relative rates in figure 2a are defined over a continuous interval from 0 to 2. The probability of y being in any subinterval of 0 to 2 is proportional to the length of the subinterval. For example, the probability of $y \leq \bar{y}$ is 0.50 because the subinterval from 0 to 1 is one half the interval from 0 to 2.

Thus, the distribution of y , as represented in figure 2a, fits the definition of a continuous uniform distribution (Haan, 1977, p. 97). The probability density function for a continuous uniform distribution is (Haan, 1977, p. 97):

$$f(x) = \frac{1}{\beta - \alpha} \quad (1)$$

where β is the largest possible value of the random variable and,
 α is the smallest value.

Thus, from equation 1, with $\alpha = 0$, it follows that the density function of relative rates is:

$$f(y) = \frac{1}{Y - 0} = \frac{1}{Y} \quad (2)$$

and the cumulative distribution function is:

$$\int_0^y f(y) dy = \int_0^y (1/2) dy = (1/2)y \quad (3)$$

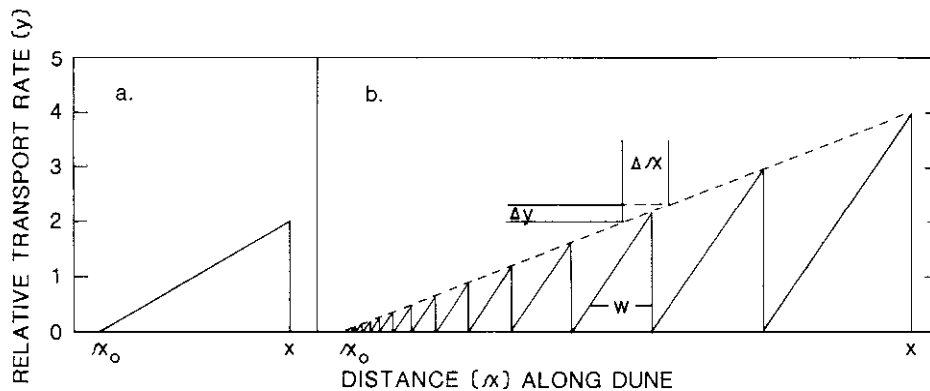


Fig. 2 Relative transport rates (y) versus distance (x) along dune for (a) dune with no secondary bedforms and (b) secondary bedforms moving along the upstream side of dune according to Hamamori's assumptions.

From observations in the flume, Hamamori concluded that the downstream movement of dunes is caused by smaller bed forms that move along the upstream slope of the dune and cascade over the crest. As these smaller, or secondary, bed forms move up the slope, they erode the primary dune and grow in height and length from zero at the upstream trough to a maximum at the dune crest. The transport rate distributions corresponding to these secondary bed forms can be represented as a series of triangles whose height increases uniformly from zero at the upstream trough to a maximum at the dune crest (fig. 2b). The mean rate corresponding to each triangular element is one half the maximum rate for that element. This mean rate represents the average transport rate for the section of the dune occupied by the corresponding secondary bed form. The overall mean transport rate is the average for the dune, which is

one half of the mean rate for the triangular element at the dune crest. Thus, transport rates vary from zero to four times the overall mean rate, and relative transport rates vary from zero to four (fig. 2b).

Hamamori's analysis results in the density function:

$$f(y) = (1/4)\ln(4/y) \quad (4)$$

and cumulative distribution function:

$$F(y) = (y/4)\{\ln(4/y) + 1\} \quad (5)$$

The distribution function is shown graphically in figure 3a. The curve shows that for the case of a dune bed and steady flow, bedload transport rates should vary from zero to four times their mean rate and 60 percent of all rates should be less than the mean. Several sets of field and laboratory data collected during nearly steady flow with dune beds correspond very closely to the theoretical distribution function (fig. 3).

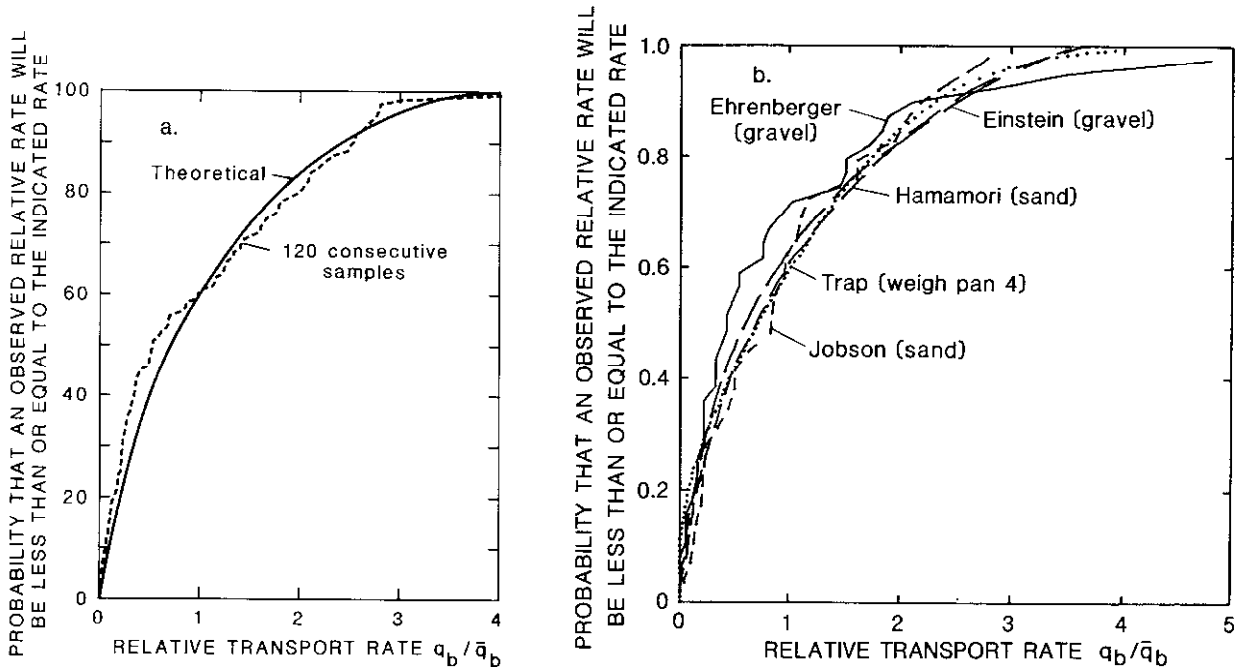


Fig. 3. Theoretical and sampled distributions of relative transport rates (from Carey, 1985 (a), and Hubbell et al., 1985 (b)).

To date, most of the available data sets that are large enough for statistical analysis have been collected during conditions for which Hamamori's derivation is applicable. However, due to the increasing incidence of bedload sampling under a variety of conditions, it is of interest to explore how Hamamori's derivation changes with different initial assumptions.

For this analysis, bedload transport is also assumed to move in oscillations or pulses. Whether or not these oscillations correlate with observable bed forms is not a consideration. The amplitudes of the oscillations are assumed to form a continuous nonuniformly increasing function from $y = 0$ at $x = 0$, to $y = Y$ at $x = X$:

$$y = ax^m \quad (m \geq 0) \quad (6)$$

where y is the peak value, or amplitude of an oscillation, and
 a is a constant for proportionality and units conversion.

The maximum relative transport rate is given by:

$$Y = 2(m+1) \quad (7)$$

The resulting generalized density function for $m \neq 1$ is:

$$f(y) = \frac{Y^{(1/m)-1} - y^{(1/m)-1}}{(1-m)Y^{1/m}} \quad (8)$$

and the generalized cumulative distribution function is:

$$F(y) = \frac{y}{(1-m)Y} - \frac{m y^{1/m}}{(1-m)Y^{1/m}} \quad (9)$$

In Hamamori's analysis, $m = 1$ in equations 6 and 7; however, equations 8 and 9 cannot be used directly when $m = 1$. By applying L'Hôpital's rule, it can be shown that Hamamori's distribution function (eq. 5), is the limit of the generalized function (eq. 9), as m approaches one.

The generalized density function, equation 8, can be used to obtain an expression for variance in terms of the value of m . The generalized expression for the variance is found to be:

$$\sigma^2 = (4m^2 + 2m + 1)/(6m + 3) \quad (10)$$

Values of Y and variance that correspond to selected values of m as computed by equations 7 and 10, respectively, are listed in table 2. Values of Y and variance for $m = 0$ are the same as those obtained for the single triangular dune analysis (eq. 2), and values of Y and variance for $m = 1$ are the same as those obtained for Hamamori's density function, equation 4 (Nijdam, 1973).

A third characteristic of the various distributions that would be useful in making comparisons is the fraction of y values that are equal to or less than the mean value. This can be found by substituting equation 7 into equation 9 and letting $y = 1$.

$$F(1) = \frac{1}{1-m} \left(\frac{1}{2m+2} - \frac{m}{(2m+2)^{1/m}} \right) \quad (11)$$

The values found from equation 11 are listed in table 2, and the changes in variance, Y, and F(1) with different values of m are shown graphically in figure 4.

Table 2. Values of Y, variance, and F(1) computed from equations 7, 10, and 11.

m	Y	Var	F(1)
0.0	2	0.333	0.50
0.5	3	0.500	0.56
1.0	4	0.778	0.60
2.0	6	1.400	0.65
3.0	8	2.048	0.69

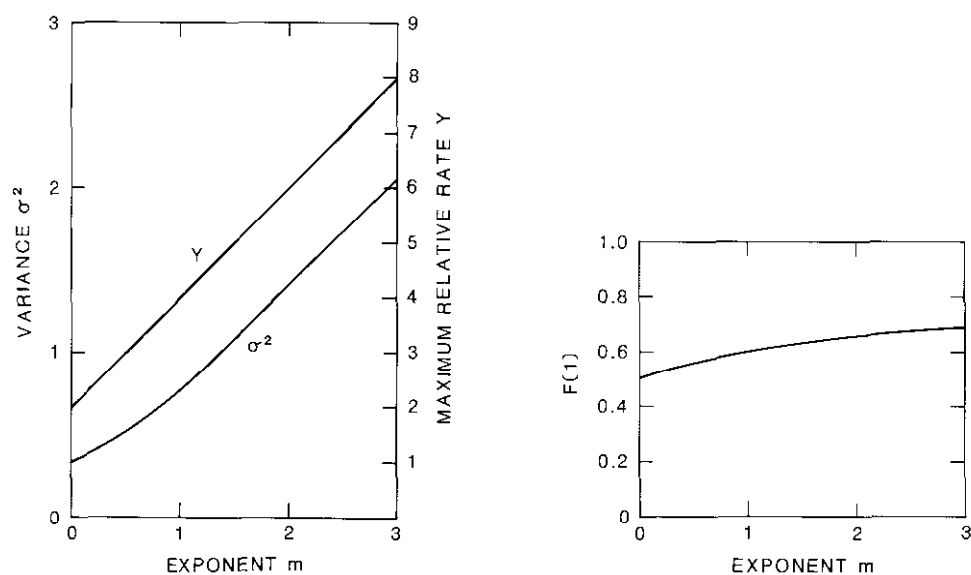


Fig. 4. Variance, Y, and F(1) for different values of m.

DISCUSSION AND APPLICATIONS

A significant aspect of the generalized functions is that they do not depend on an observed relation between transport rate oscillations and specific bed-form geometry. Thus, for cases where transport rate oscillations may exist without detectable bed forms, the generalized functions may provide descriptions of the distribution of transport rates once a value for m has been determined from theory or experiment. However, since measured data are lacking to illustrate applications of the generalized functions, it is

convenient to use the bed-form analogy in studying possible applications. Since Hamamori's assumption of uniformly increasing peak rates is necessary for maintaining constant dune shape, values of $m \neq 1$ in the general derivation imply that dune shape must be changing.

Predicting the unsteady behavior of dunes is complicated by the fact that both shape and velocity may change with time. Klaassen (1979) has shown that changes in dune height occur much faster than changes in length because changes in height are related to changes in individual dunes, whereas changes in length are related to interactions among different dunes. If changes in velocity are assumed to be negligible, it is possible to show that values of m greater than 1 may represent conditions where dune height is decreasing and m values less than 1 may represent conditions where dune height is increasing. Although these conditions of bed-form adjustment are transient, it has been shown that modifications to bed forms occur slowly even when discharge is changing rapidly (Fredsoe, 1979). While it is unlikely that a particular distribution applies continuously during transient conditions, the adjustment of bed forms occurs slowly enough that trends illustrated in figure 4 can be used to make some preliminary inferences about expected variability. For $m > 1$, both Y and variance increase rapidly and $F(1)$ increases gradually; whereas for $m < 1$, opposite trends are indicated (fig. 4). Thus, when dune heights are decreasing, the variance of bedload samples collected at a fixed point may be larger than the variance during steady-state conditions, and when dune heights are increasing, the variance may be less. If this is true, it would indicate that to achieve a given level of accuracy fewer samples would be required when dune heights are increasing than when steady-state conditions are present. Verification of this possibility and definition of applicable m values would affect recommendations for bedload-sampling procedures during unsteady conditions.

One of the most frequent applications for quantitative descriptions of hydrologic phenomenon is the design of efficient sampling procedures. An efficient procedure can be defined as one that properly accounts for the inherent variability of the process being sampled while keeping the time spent and the number of samples collected to a minimum. As with most hydrologic sampling efforts, the degree of accuracy that is required depends on the objectives of the sampling program. Thus, by knowing the accuracy of different procedures, the optimal procedure can be selected.

A common objective of bedload sampling is to determine the distribution of mean rates in the cross section. Given the scale of inherent variability, it is obvious that numerous samples have to be collected at each cross-section location. Statistical sampling theory states that:

$$\sigma(\bar{y}) = \frac{\sigma(y)}{\sqrt{N}} \quad (12)$$

where $\sigma(\bar{y})$ and $\sigma(y)$ are respectively, the standard deviation of the means of sets of N samples, and the standard deviation of individual samples. Equation 12 is plotted as a family of curves with each curve corresponding to a selected value of m (fig. 5). The curves are based on standard deviations obtained from the variances in figure 4. These curves can be interpreted as indicating that the mean determined from N samples will be within $\pm\sigma(\bar{y})(100)$ percent of the true mean at the sampling point about 68 percent of the time. Other confidence intervals can be obtained from the t -distribution. Figure 5

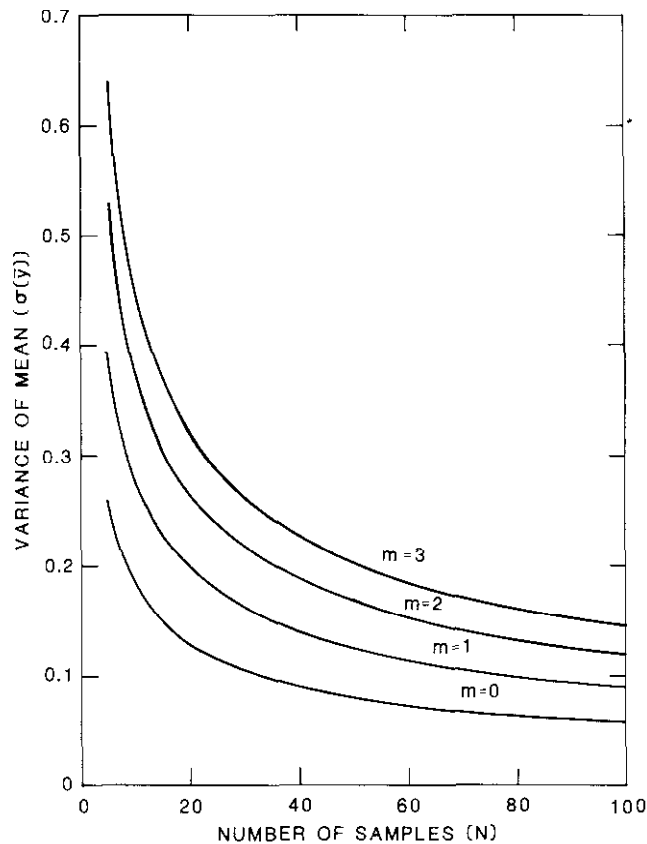


Fig. 5. Variance of mean versus numbers of samples for selected values of m .

shows that for any of the curves a mean rate determined from fewer than 5 samples has little value.

Interpretations drawn from figure 5 are based on the assumption that a perfect bedload sampler is being used and that sampling is random. Hubbell et al., (1985) discuss tests with the Helley-Smith bedload sampler (Helley and Smith, 1971) and present calibration curves for adjusting sampled rates to represent true rates. When the sampling location is fixed, for example on a bridge, then random sampling of bedload transport is very difficult; however, sequential sampling at a point is relatively easy. Sequential samples of bedload transport are highly serially correlated, and must be converted to equivalent numbers of random samples before figure 5 can be used. Yevjevich (1972, p. 187-188) discusses some methods of making this conversion.

SUMMARY

In order to obtain accurate samples of bedload transport, large-scale inherent variability must be accounted for in the sampling procedure. Several sets of field and laboratory data have shown that an existing probability density function and corresponding distribution function can be used to describe this inherent variability for conditions of nearly steady flow over a dune bed. With the current popularity of bedload sampling, many sampling efforts are being undertaken in conditions for which the existing functions may not apply.

In order to provide more generally applicable sampling guidelines, a density function and corresponding distribution function have been derived that do not depend on a specific bed-form geometry. A general expression for the variance has also been derived, and used to illustrate how sampling requirements change for specific forms of the density function. These specific forms of the generalized functions have not been verified by actual data, because detailed data sets are only available for the case of nearly steady flow with distinct bed forms. Possible applications for the generalized functions include, conditions where dune height is changing, and conditions where bed forms are not distinct.

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BED FORM PREDICTORS FOR HIGH SEDIMENT LOADS

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ABSTRACT

During the Mt. St. Helens eruption 18 May 1980, a debris avalanche deposited some two and one-half billion cubic meters of rock, ice, and other material in the upper twenty-seven kilometers of the North Fork Toutle River Valley. Mud flows triggered by the eruption carried large volumes of sediment from the debris avalanche into the Toutle, Cowlitz, and Columbia Rivers. Three years after the initial eruption, large volumes of sediment continue to be delivered to these rivers from the debris avalanche. High wash load (less than 0.062 mm) concentrations have been observed to significantly affect streamflow characteristics and bed material transport (sand) in the river system. As the wash load concentration increases, fluid viscosity increases, the sediment particle fall velocity decreases, and bed material transport increases. Variations in wash load concentration alters bed forms which in turn influence peak flows and travel times from the upper watershed, and water surface elevations in the Cowlitz River. Water surface profiles have varied as much as one meter in the Cowlitz when dunes have transitioned to a plane bed as a result of high wash load concentrations. Four bed form predictors; Simons and Richardson, two by Athallah, and Vanoni, are compared in this paper. The Vanoni relationship best fitted observed field data. An approach is also presented which incorporates sediment concentration as these methods do not include it as a variable.

INTRODUCTION

On 18 May 1980, just before a magmatic blast signalled the onset of the Mt. St. Helens eruption, an earthquake triggered a major slope failure on the north flank of the mountain. As a result, approximately two and one-half billion cubic meters of rock and ice avalanched into the North Fork Toutle River Valley. Mud flows and debris flows were generated and swept down the Toutle and Cowlitz Rivers destroying bridges, inundating buildings, and eventually blocking the navigation channel of the Columbia River. Within twenty-four hours the mudflows deposited some forty-six million cubic meters of sediment in the lower thirty-seven kilometers of the Cowlitz River reducing channel capacity by nearly eighty percent. The areal extent of the eruption impact is shown in Figure 1.

By mid-May 1981, a year after the eruption, the Cowlitz River channel had been restored to roughly pre-eruption capacity, as nearly 51.7 million cubic meters had been dredged from the river by the Corps of Engineers [U.S. Army Corps of Engineers, 1982]. Since the completion of dredging, the Cowlitz has aggraded markedly as a result of tremendous sediment yields from the debris avalanche in the North Fork Toutle River. Variations in streambed slope, channel geometry and river hydraulics have been dramatic in the upper Cowlitz and Toutle Rivers.

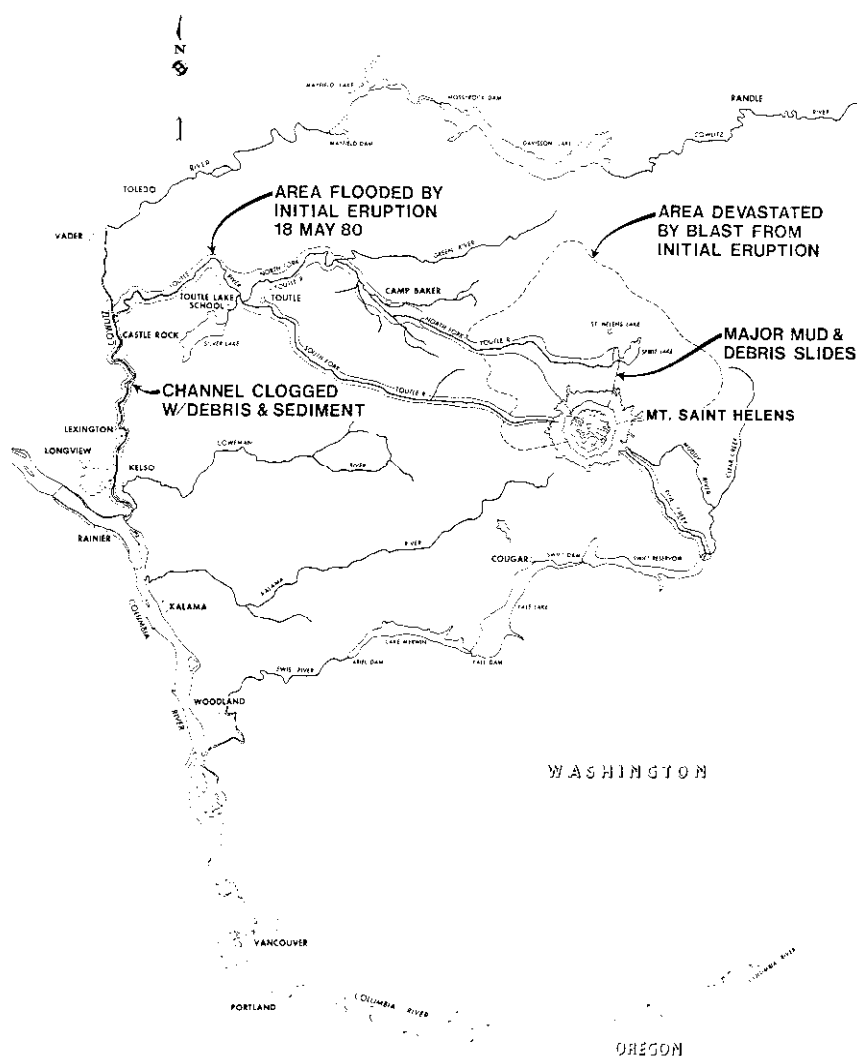


Fig. 1. Area map.

meter for similar streamflows depending on the type of bed form prevalent in the Cowlitz River. This has a major impact on estimates of flood protection levels [U.S. Army Corps of Engineers, 1982].

Due to the importance and variability of bed forms in this river system, the effectiveness of predictive relationships was assessed at three gaging stations; the Cowlitz River at Castle Rock, the Toutle River at Highway 99 Bridge, and the Toutle River at Tower Road. The bed form predictors evaluated include those by Simons and Richardson [1966], two by Athallah [1968], and Vanoni [1981]. These relations do not include sediment concentration as a variable, yet it has been shown to be extremely important for this river system [Graham and Bradley, 1982; U.S. Army Corps of Engineers, 1982]. It's importance to any effective bed form predictor was assessed and future data needs addressed to develop an appropriate formulation including sediment concentration.

High sediment delivery to the Toutle and Cowlitz Rivers has significantly changed channel roughness and bed form since the 1980 eruption. The Toutle River was a pool-riffle stream under pre-eruption conditions composed of bed material varying from gravel to boulders. It became a sand and gravel bed river with predominantly upper regime bed forms, increased flood peaks, and decreased travel times following the eruption. The Cowlitz River was a gravel bed stream prior to the eruption, which has been transformed to a sand bed stream as a result of extremely high transport of sands and silts from the debris avalanche. Depending on wash load concentrations, bed forms change from dunes to plane bed and antidunes. Observed water surface elevations have varied as much as one

CHANGES IN SEDIMENT YIELDS AND TRANSPORT RATES

The initial eruption of Mt. St. Helens created both long- and short-term impacts on the sediment transport processes in the Toutle and Cowlitz Rivers. A massive new sediment source, non-existent prior to the eruption was added to the system. Erosion processes in the debris avalanche deposits have been changing constantly during the past few years. Sediment delivery to the Toutle is dependent upon many factors, including basin hydrology and individual sub-basin contribution, the state of new channel development in the avalanche, antecedent moisture conditions in the avalanche which lead to mass wasting and bank failure, and mud flows generated by volcanic events. Once sediment is delivered to the rivers, transport rates are dictated by channel geometry, bed slope, bed material sediment size, and wash load sediment concentration. Variations in sediment transport can be seen in a comparison of measured suspended sediment transport rates as shown in Figure 2, for pre-eruption conditions, the January 1982 storm, the February 1982 storm, and the

19 March 1982 mud flow at the Toutle River Highway 99 Bridge gaging station. This gage is one mile above the Toutle-Cowlitz confluence and provides a measure of the inflowing sediment loads to the Cowlitz River.

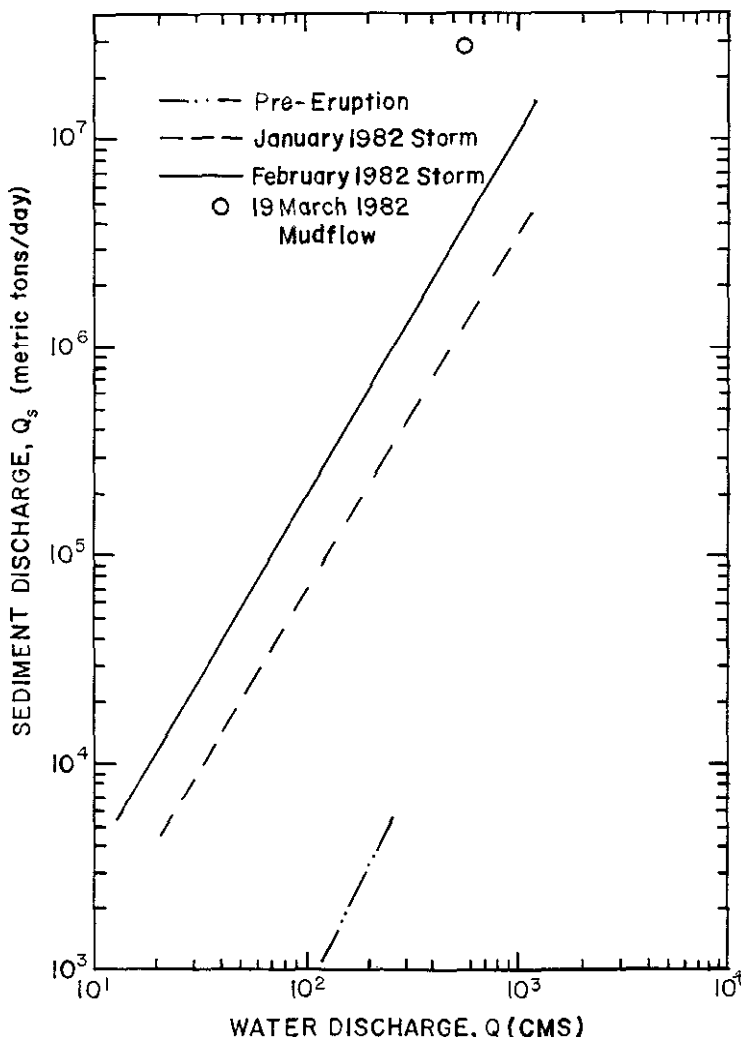


Fig. 2. A comparison of measured suspended sediment transport rates.

As a result of the new sediment source, Toutle River sediment transport rates have increased approximately two to three orders of magnitude since the Mt. St. Helens eruption. Sediment transport rates were approximately half an order of magnitude greater in the February storm event than in January at similar streamflows for the following reasons: the North Fork Toutle River sub-basin runoff and sediment yield was greater in February; mass wasting and bank failures due to high antecedent moisture content were followed by local mud flows in the debris avalanche in February 1982; and high wash load or fine sediment concentrations (less than 0.062 mm) kept bed material (sand) in transport for greater periods of time in the February storm [Graham and

Bradley, 1982]. On 19 March 1982 an explosive eruption in the volcano's crater melted a large quantity of snow that generated a debris flow and mud flow which moved rapidly downstream into the Cowlitz River. This event produced sediment transport rates at the Toutle River Highway 99 gaging station almost an order of magnitude higher than those observed in the February 1982 storm event.

Sediment yields from the debris avalanche deposits will continue to be extremely high and variable for years to come, whether the transport mechanism is a mud flow or more normal transport phenomena. Extensive channel adjustment will continue in the Cowlitz and Toutle Rivers, as channel slope, width, depth, and roughness adjust to high magnitude sediment loads.

CHANNEL ADJUSTMENTS

The lower 31 kilometers of the Cowlitz River aggraded approximately 7 meters due to the initial eruption. Bed slope began increasing due to high sediment yields from the Toutle River as soon as the channel dredging of 1980 and 1981 ended. Initially, the channel was overdredged above kilometer 26 actually reducing the bed slope from that of pre-eruption conditions. The channel had aggraded to approximately pre-eruption levels by 5 February 1982, and almost to post-eruption channel conditions, as they existed following the initial eruption, by 5 March 1983. The Corps of Engineers has continued to dredge the Cowlitz River to restore channel flood capacity since the summer of 1983. This is a stopgap preventative measure and even though bed slope has been reduced temporarily, it will again increase so long as sediment yield to the reach remains high.

As streambed slope and streambed elevations increased and bed material size or particle roughness decreased after the eruption, channel widths have increased and flow depths have decreased in the Toutle and Cowlitz Rivers. Surveyed cross sections on the North Fork and Main Fork of the Toutle River indicate that channel widths increased 200 to 300 percent during the 1982 water year. The Cowlitz River channel is also widening. Width-depth ratio and velocity versus discharge relationships during three storm events at the Cowlitz River at Castle Rock gaging station are shown in Figure 3. Streamflow hydrographs were similar at this gage for the storms of January 1982, February 1982, and December 1982. Differences between these relations indicate the extensive variability of system response to different conditions. Greater velocities and smaller flow depths in February as compared with January were the result of different bed forms. Based on field observations, it is known that dune bed forms produced approximately one meter greater flow depths in January than the plane bed observed in February. This effect was attributable to high wash load concentrations during the February storm event [Graham and Bradley, 1982; U.S. Army Corps of Engineers, 1982]. Hydraulic relationships were very similar between the February and December storms. This similarity was not due to high wash load concentrations in December, but was a result of finer bed material sizes and increased bed slope during that storm.

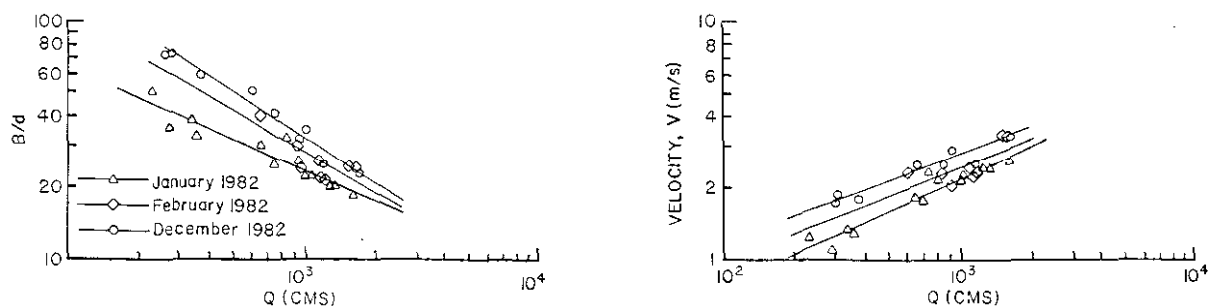


Fig. 3. Width - depth ratio versus discharge, and velocity versus discharge Cowlitz River at Castle Rock.

BED FORM PREDICTIONS

Predictions and Comparisons With Observed Prototype Data

The type of bed forms found in the Toutle and Cowlitz River system impact peak flows and travel times in the Toutle River, and water surface elevations in the Cowlitz. Observed hydraulic parameters and sediment properties were used to evaluate bed form predictors at the three gaging stations; the Cowlitz River at Castle Rock, the Toutle River at Highway 99, and the Toutle River at Tower Road. Bed form predictors evaluated include those of Simons and Richardson [1966], two by Athaullah [1968], and Vanoni [1981]. A comparison of observed bed forms and those predicted by these methods are summarized in Table 1, and in the following remarks.

The relationship developed by Simons and Richardson [1966] is compared with observed bed form only at Castle Rock, as bed slope data needed for the analysis are unavailable at the Toutle River stations. This predictor relates streampower to median fall diameter of the bed sediment. As shown in Figure 4, bed forms for the January 1982, February 1982, and December 1982 storm events were predicted to be upper regime. Field observations of bed forms are difficult to make in the Cowlitz River due to large flow depth and high turbidity levels, but it has been observed that dunes occur at both high and low flows and become transitional or plane bed at high flows with high wash load concentrations. This was the case observed by USGS and Corps of Engineers field personnel during the January and February 1982 storms [U.S. Army Corps of Engineers, 1982]. A plane bed existed at high flows (1400-2000 cms) with high wash load concentration in February, while dunes were observed in January for similar streamflow but with much lower wash load concentrations. During December 1982, a plane bed was observed at similar flows as those of January and February, yet it was the result of increased bed slope and decreased bed material size, not high wash load concentrations. This bed form predictor over-estimates the percent of upper regime conditions for the Cowlitz River when compared to field observations which indicate a high percentage of dunes, particularly for lower flows as seen in all storm events.

Table 1. Observed versus predicted bed forms.

GAGING STATION	STORM EVENT	OBSERVED BED FORMS	SIMONS & RICHARDSON	ATHAULLAH S vs $1/\psi$	ATHAULLAH F vs d/D_{50}	VANONI
Cowlitz River at Castle Rock	January 1982	Dunes	Upper Regime (Antidunes and Plane Bed)	Transition	Transition	Dunes
	February 1982	Plane Bed at High Flow Dunes at Lower Flows	Upper Regime (Antidunes and Plane Bed)	Antidunes	Upper Regime (Antidunes and Plane Bed)	Flat Bed
	December 1982	Plane Bed at High Flow Dunes at Lower Flows	Upper Regime (Antidunes and Plane Bed)	Transition	Upper Regime (Antidunes and Plane Bed)	Flat Bed
Toutle River at Highway 99	January 1982	Plane Bed and Antidunes	-----	-----	Transition	Flat Bed
	February 1982	Plane Bed and Antidunes	-----	-----	Transition	Flat Bed
Toutle River at Tower Road	January 1982	Plane Bed and Antidunes	-----	-----	Transition and Upper Regime (Antidunes and Plane Bed)	Flat Bed and Anti- dunes
	February 1982	Plane Bed and Antidunes	-----	-----	Transition and Upper Regime (Antidunes and Plane Bed)	Flat Bed and Anti- dunes

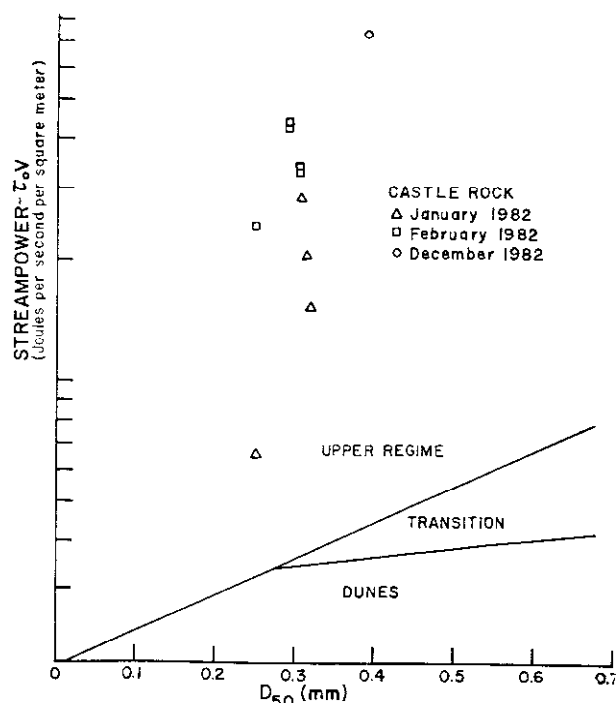


Fig. 4. Comparison of observed bed forms with Simon's and Richardson's predictor.

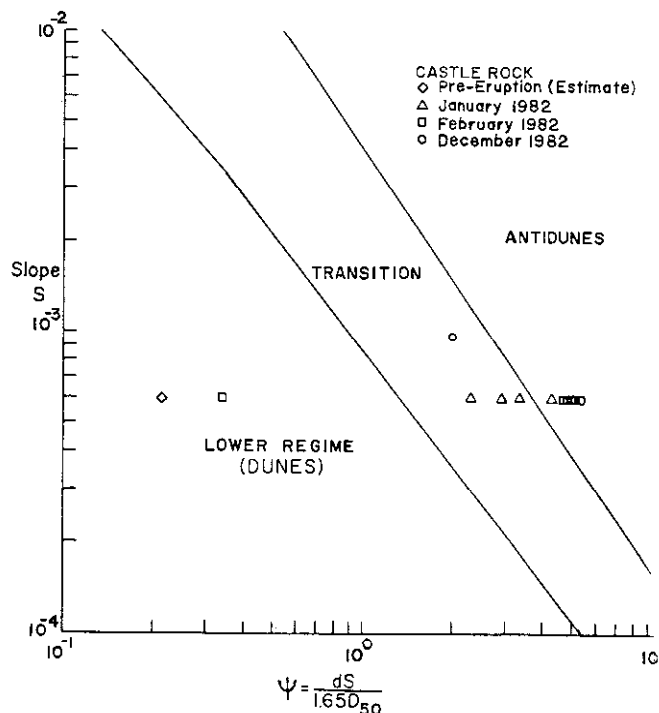


Fig. 5. Comparison of observed bed forms with Athaullah's predictor using slope and shear intensity factor.

A bed form predictor developed by Athaullah [1968] seems to better compare with field observations at Castle Rock. Data shown in Figure 5 for the same three storm events plot within the transition or antidune range. However, as with the previous method this procedure does over-estimate the percentage of upper regime occurrences versus dune forms as compared to the prototype. This is a bed slope relation and can only be applied at Castle Rock where adequate slope information is available.

Another procedure developed by Athaullah [1968] relates Froude Number to relative roughness, d/D_{50} . Hydraulic and sedimentological data from all three gages are used to forecast bed forms as plotted in Figure 6. Bed forms are

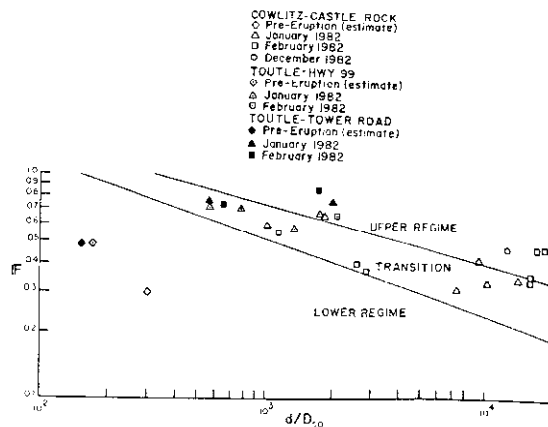


Fig. 6. Comparison of observed bed forms with Athaullah's relationship using Froude Number and relative roughness.

estimated to be transitional or upper regime at all three gages which is certainly in error at Castle Rock for lower flow rates. The Toutle River predictions are much more reasonable based on field observations. The dominant bed form pattern observed for the entire flow range in the Toutle River is presented in Figure 7, showing antidunes at low to moderate discharge. This predictor does well for Toutle River stations where bed forms are almost always transitional or upper regime, but again over-estimates these conditions at Castle Rock.



Fig. 7. Observed bed forms in the Toutle River immediately upstream of the Toutle-Cowlitz confluence.

Vanoni's [1981] method, which also plots Froude Number versus relative roughness, seems to be the best of the four relationships for these three gaging stations. Figure 8 shows most Castle Rock predictions to plot as dunes in January and plane or flat bed during the February event. These results compare favorably with observed Cowlitz River bed forms. Toutle River gage data plots mostly in the flat bed zone with a few points indicated as antidunes. This may be a slight under-estimate of antidunes in the Toutle River but is well within the range of field observations. Figures 6 and 8 both include for reference an estimate of pre-eruption bed form patterns at the gaging stations. Comparison of pre-eruption and present conditions portrays the dramatic change in bed forms since the eruption.

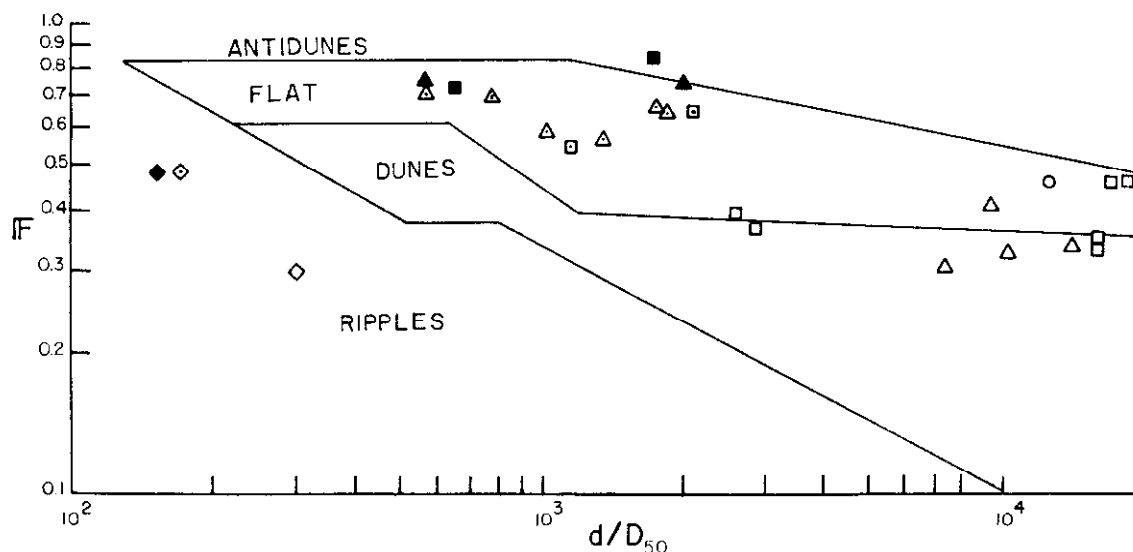


Fig. 8. Comparison of observed bed forms with Vanoni's relation.

A major shortcoming in the preceding evaluation is its basis on minimal field data, which is of both a qualitative and quantitative nature. It was observed by field personnel that dunes did shift to a plane bed at peak flows in February but not January. This was verified in standard step-backwater computations (HEC-2) of this river reach. Manning's n , channel roughness values, had to be reduced due to variations in form roughness from 0.03 in January to 0.019 in February to match observed water surface elevations. These observations and backwater analyses were all made for peak streamflows. Most of the observed hydraulic and sediment data used to predict bed forms are, however, for lesser flows and sediment concentrations. Maximum suspended sediment concentrations observed in the Cowlitz in January were approximately 20,000 ppm as compared to 100,000 ppm in February. This lack of high flow and concentration data is of import in Cowlitz River predictions as the bed forms change from dunes to plane bed under these conditions. Toutle River estimates are not as critical since transitional and upper regime conditions exist at even low to moderate flows. Perhaps the most important problem in using these prediction techniques for this river system is that they do not include sediment concentration as a variable.

Predictions Incorporating Sediment Concentration

The effects of sediment concentration on bed forms in the Toutle and Cowlitz Rivers are portrayed in Figure 9. A plot of Froude Number versus relative roughness is shown. Measured suspended sediment concentrations, C_t , and wash load concentrations, C_f , are also tabulated. No pattern can be seen in the Toutle River data showing sediment concentration to be a major variable in determining bed form. This may be true simply because upper regime bed forms are observed in most situations at these sites even for low discharges. A pattern is noted in the Cowlitz River data, as it appears that bed forms will shift towards a higher regime (i.e., plane bed for the case shown) both as a function of variable hydraulic parameters and increasing concentration. It is necessary to add to the prototype data base at higher flows and concentrations to verify the magnitude of this effect.

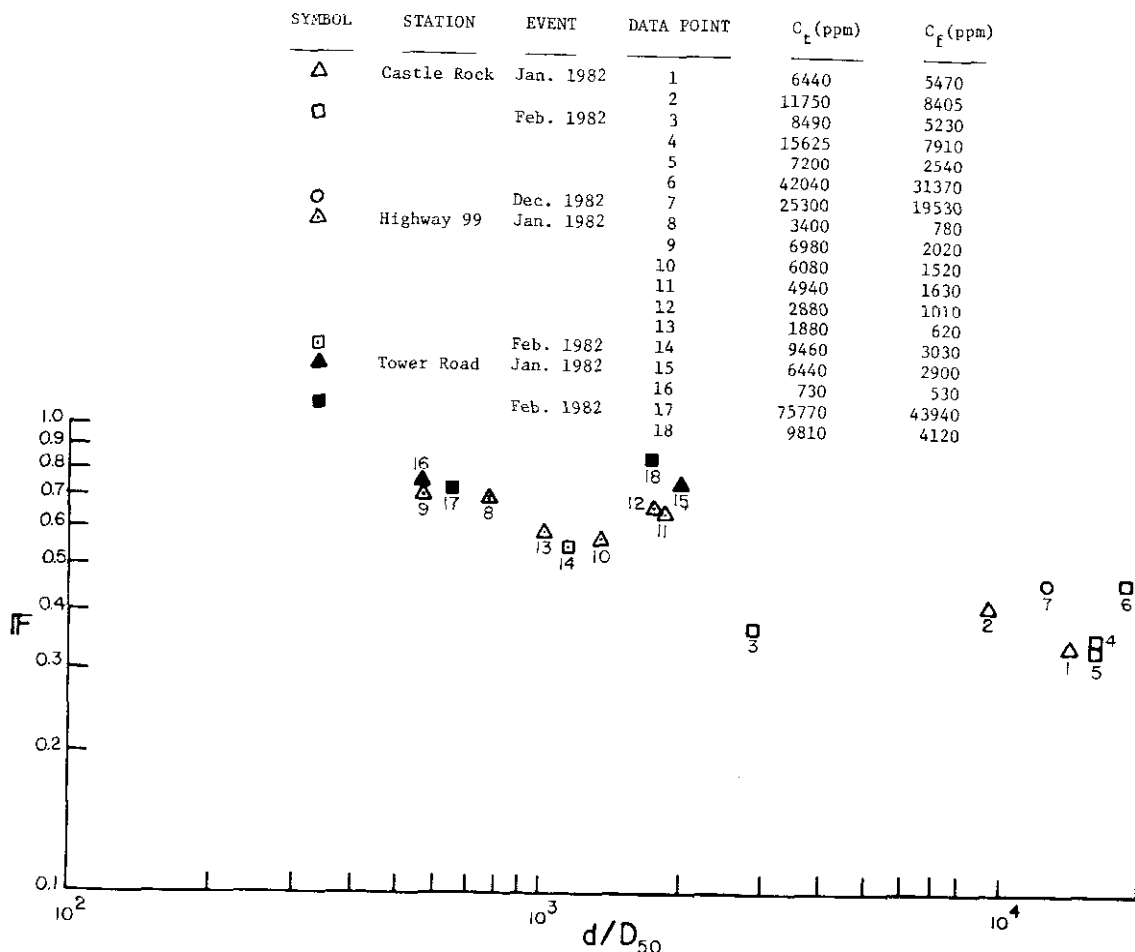


Fig. 9. Comparison of Froude Number versus relative roughness as a function of suspended and wash load concentrations.

Fluid properties have been noted to vary significantly with sediment concentration as has been documented by Thomas [1963]. Bed forms are in turn affected by changes in concentration and fluid properties. Based on Thomas' [1963] approach, resultant decreases in particle fall velocity, w , and increases in viscosity, μ , with increasing concentration are shown in Table 2 for the Cowlitz River. Fall velocity is seen to decrease as much as 21 percent and viscosity to increase 18 percent of observed concentrations.

In pioneering work done in bed form prediction using dimensionless plots Liu [1958] related V_*D/ν and V_*/w in defining the line of demarcation for ripple formation, where V_* is the shear velocity. Albertson, Simons and Richardson [1958] used these same parameters to evaluate higher order bed forms in flumes. Unfortunately, verification of their criterion by comparison with river data demonstrated the inaccuracy of using this predictor for rivers. As the criteria is based on flume data with steep slopes and low depths it is not valid for application to alluvial streams with mild slopes and large depths. This method if extended to cover alluvial channels for prototype data may be applicable for the Cowlitz River as it does incorporate the effects of sediment concentration in the fall velocity and viscosity terms. Future data collection should document the magnitude of the effect of sediment concentration on bed form, and allow the development of a more

Table 2. Variation of fluid properties with sediment concentration.

STATION	DATE	C_T (ppm)	$\frac{w_s}{w_o}$	$\frac{\mu_s}{\mu_o}$
Cowlitz River at Castle Rock	January 23, 1983	6,400	0.99	1.07
	January 15, 1983	11,750	0.97	1.07
	Maximum Observed in January Storm	20,000	0.96	1.08
	February 16, 1983	8,490	0.98	1.07
	February 17, 1983	15,625	0.97	1.08
	February 19, 1983	7,200	0.98	1.07
	February 20, 1983	42,040	0.91	1.11
	Maximum Observed in February Storm	100,000	0.79	1.18
	December 4, 1983	25,300	0.94	1.09

* Based on Thomas Equation

$$\frac{\mu_s}{\mu_o} = 1 + 2.5 \phi + 10.05 \phi^2 + 0.062 \exp. \left(\frac{1.875 \phi}{1-1.595 \phi} \right)$$

$$\ln \frac{w_s}{w_o} = -5.9 \phi$$

μ_o, w_o - clear water properties

μ_s, w_s - sediment laden fluid properties

adequate bed form predictor using the aforementioned approach or a facsimile thereof.

CONCLUSIONS

Channel adjustments and sediment concentrations have been extremely large since the initial eruption of Mt. St. Helens. The general pattern of channel change is one of increasing slope, decreasing flow depth, and increased width. Suspended sediment and wash load concentrations have been observed to vary as much as an order of magnitude due to a number of factors. High wash load concentrations significantly control the type of bed form in the river system. This variation in bed form has caused significant fluctuation in water surface elevations in

the Cowlitz River, and must be taken into account in any adequate evaluation of flood risk in the Cowlitz River floodplain.

Four bed form predictors were evaluated, with the Vanoni [1981] procedure comparing most favorably with field observations. None of the methods include sediment concentration as a variable. Use of this method is adequate in the Toutle River where upper regime conditions predominate, but is questionable for the Cowlitz River due to major differences in bed form with variation in wash load concentration. A predictor is needed for the Cowlitz which incorporates sediment concentration as a major variable. A method similar to that originally formulated by Liu [1958] and extended by Albertson, Simons, and Richardson [1958] which does include the effects of concentration is recommended for use in the Cowlitz River. Additional field data are essential, however, in order to formulate a substantive procedure to predict bed forms and channel roughness where sediment concentration is highly variable.

ACKNOWLEDGMENTS

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SUSPENDED-SEDIMENT TRANSPORT, LAKE TAHOE BASIN

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ABSTRACT

Analysis of magnitude and frequency of suspended-sediment transport in four tributary streams of Lake Tahoe indicated that from 1973 to 1981 suspended sediment was transported primarily by frequent, moderate streamflow. Spring snowmelt transported 73.8 percent of the total annual suspended-sediment load. Ninety percent of the suspended sediment in these four streams was transported by flows that occurred more than 1.4 days per year. Less than 13 percent of the total suspended-sediment load was transported by flows that occurred only 1 percent of the time.

The relationship of suspended-sediment discharge to water discharge was found to depend upon factors such as amount of snow cover and the type of runoff that produced streamflow. Data were divided into four groups according to the time of year during which different runoff processes prevail and were further subdivided into high and low discharge classes within each seasonal group.

INTRODUCTION

Lake Tahoe is famous for the extreme clarity and purity of its water. During the last 25 years, however, increases in rates of primary productivity and sediment deposition in the lake have created concern about its water quality (Byron and Goldman, 1984). Because the delivery of nutrients to the lake is related to the transport of suspended sediment (Goldman, 1974), an understanding of sediment-transport processes is needed to properly manage tributary drainage basins. The objectives of this study are to analyze the magnitude and frequency of suspended-sediment transport by selected streams in the Lake Tahoe basin, and to determine the streamflow conditions most responsible for delivering suspended sediment to the lake.

In addition to their relationship to nutrient transport, the magnitude and frequency of sediment transport by streams can be used to evaluate the amount of "geomorphic work" accomplished by various types of streamflow. Wolman and Miller (1960) discussed, in terms of geomorphic work, the ability of geomorphic processes to transport geologic materials and to sculpt the landscape. They noted that the frequency of occurrence of a particular event, as well as its magnitude, determined the significance of the event in the formation of the landscape. Information collected on Brandywine Creek, Delaware, and on Rio Puerco, New Mexico, indicated that, although extreme flows with recurrence intervals greater than 50 to 100 years were capable of transporting large quantities of sediment, the relatively infrequent occurrence of such flows reduced their geomorphic effectiveness. Moderate, relatively frequent flows were found to be more effective in shaping the landscape.

LAKE TAHOE BASIN

Physical Setting

The Lake Tahoe basin is located in the northern Sierra Nevada (fig. 1). Lake Tahoe is oval in shape, about 22 miles long from north to south and about 12 miles wide; the Lake Tahoe basin, also oval, is approximately 40 miles long and 10 to 18 miles wide (Crippen and Pavelka, 1970). Lake Tahoe occupies a graben in the Sierra Nevada formed during the late Tertiary Period. The altitude of the lake surface has been raised to 6,229 feet above sea level by a small dam.

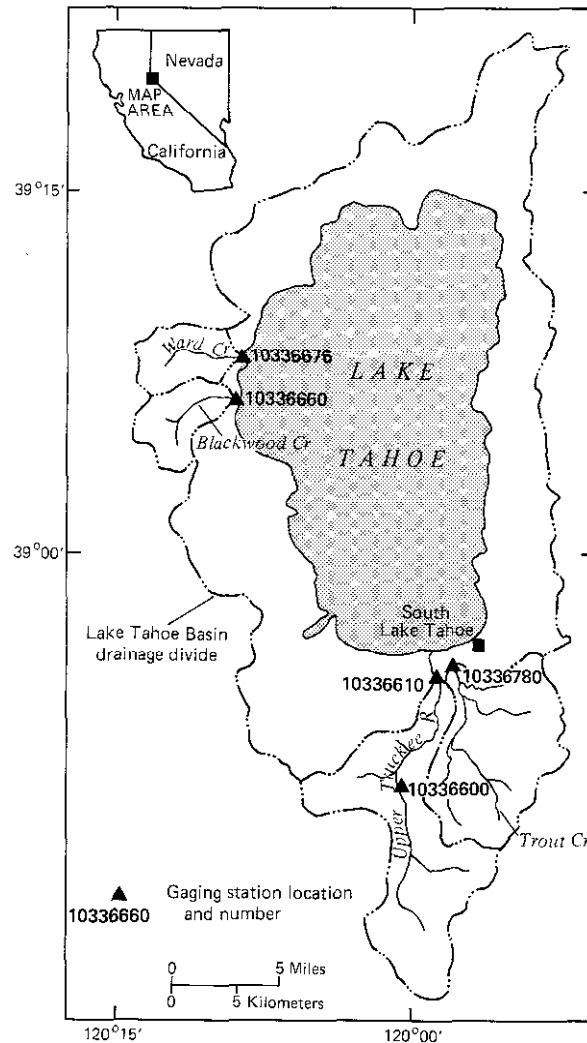


Fig. 1. Map showing location of study area.

Geology

The region is underlain by igneous and metamorphic rocks. About 50 percent of the basin area, including much of the eastern and southern areas, consists of Jurassic and Cretaceous granitic rocks. Younger volcanic rocks occur extensively in the northwest part of the basin and to a smaller degree in the southern part (Crippen and Pavelka, 1970). Metamorphic rocks form outcrops in the southeast and southwest parts of the basin (Tahoe Regional Planning Agency, 1971). The western side of the basin was glaciated during the Pleistocene Epoch. Glaciers scoured most areas down to bedrock and deposited moraines through which streams flow. Most of the eastern side of the basin remained free from glacial ice. Consequently, the bedrock, mostly granitic, is deeply weathered and more easily eroded. Slopes in the basin range from nearly level to extremely steep. Twenty-two percent of the land surface has slopes of less than 10 percent; 50 percent of the land area has slopes greater than 20 percent (Crippen and Pavelka, 1970).

Climate

The climate of the region is dominated by the weather patterns of the Pacific Northwest combined with orographic effects created by the Sierra Nevada. Precipitation is strongly seasonal, occurring mostly as snow during the winter months. Occasional thundershowers contribute some rainfall during the summer. Mean annual precipitation ranges from 15 to 90 inches, with precipitation greatest at high altitudes on the west side of the lake and decreasing to the east. Temperatures range from a mean daily minimum of 17°F in January to a mean daily maximum of 78°F in July.

Drainage Basins Chosen for Study

Drainage basins were selected for this study on the basis of the length and completeness of water- and sediment-discharge records (table 1). The longest suspended-sediment-discharge records were used. Locations of the study drainage basins are shown in figure 1. Drainage areas, shown in table 1, were obtained from Jorgensen and others (1978). Records of mean daily water and suspended-sediment discharge were collected and published by the U.S. Geological Survey (1974-75a, 1974-75b, 1976-82). Data from the Upper Truckee River at Meyers (10336600) were used for periods when data for the South Lake Tahoe gage (10336610) did not exist (fig. 1).

Streamflow in Study Streams

The streamflow regime in the region is dominated by late spring snowmelt, with low base flows during all but the third quarter of the water year (April through June). The study period primarily contains years with normal runoff, except for the drought years of 1976 and 1977 when flows were extremely low. Wetter than normal years occurred in 1974 and 1980. During the study period, an estimated 5- to 10-year flood occurred on Blackwood Creek and an estimated 10- to 25-year flood occurred on Ward Creek. The remaining annual maximum mean daily discharges have recurrence intervals of less than 5 years.

Table 1. Suspended-sediment transport equations for large water discharges for study.

[Equations are of the form: $\log (T + 1) = a + b (\log (Q + 1))$]

Drainage area (mi ²)	Mean daily suspended-sediment record during study period	Quarter	Water discharge to which equation applies (ft ³ /s)	Transport equation parameters (a) (b)		F-values			
						Equality of means		Equality of slopes	
						Value	Probability	Value	Probability
Upper Truckee River at South Lake Tahoe (10336610)									
54.80	10/72-6/74;10/77 6/78; 3/80-9/81	1	≥38.4	-2.44	1.64	15.7	0.00	22.9	0.00
		2	all values	-1.70	1.20				
		3	≥79.2	-3.38	1.93				
		4	≥43.5	-2.46	1.56				
Blackwood Creek near Tahoe City (10336660)									
11.16	10/74-9/76;10/77- 6/78;10/79-9/81	1	≥17.4	-2.08	1.66	9.5	0.00	19.9	0.00
		2	≥39.8	-4.17	2.62				
		3	≥44.0	-3.57	2.24				
		4	≥32.2	-1.43	0.99				
Ward Creek at Highway 89 near Tahoe Pines (10336676)									
9.70	10/72-11/76;1/77- 6/77;10/77-6/78; 10/79-9/81	1	≥22.5	-3.19	2.36	37.7	0.00	4.16	0.00
		2	≥40.8	-3.77	2.37				
		3	≥52.4	-3.63	2.21				
		4	≥45.2	-3.59	2.19				
Trout Creek near Tahoe Valley (10336780)									
36.70	10/73-9/74;10/77- 6/78;3/80-9/81	1	≥19.2	-2.38	1.89	29.0	0.00	8.06	0.00
		2	≥18.9	-1.69	1.40				
		3	≥28.8	-2.54	1.82				
		4	20.4≤Q<81.2 ≥81.2	-1.32 -1.24	1.06 1.22				

FLOW-DURATION ANALYSIS

Flow-duration analyses provided the distribution of streamflow frequency which, combined with sediment-transport data, was used to calculate the magnitude and frequency of suspended-sediment transport (Miller, 1951). Flow-duration analyses were performed using the daily values statistics program of the U.S. Geological Survey. The flow-duration analyses were done first for the entire period of record. Water-discharge information then was subdivided according to quarters of the water year (October-December, January-March, April-June, July-September) and flow-duration analyses done for each quarter separately for the entire period of record. The water year was divided into quarters because different hydrologic conditions and sediment-delivery processes seem to prevail at different times of the year (Walling, 1974, 1977a). During the first quarter (October-December), precipitation falls mostly as snow on varying amounts of bare ground. During the second quarter (January-March), precipitation occurs primarily as snow on snow-covered ground. During the third quarter (April-June), snowmelt runoff predominates and the amount of snow-covered ground decreases. During the fourth quarter (July-September), precipitation occurs as thundershowers on bare ground. During quarters 1 through 3, rain on snow-covered ground is also possible (Galton, 1984).

SUSPENDED-SEDIMENT TRANSPORT RELATIONSHIPS

The rate of transport of suspended sediment for the four streams was related to streamflow by a linear logarithmic equation of the form

$$\log(T + 1) = a + b[\log(Q + 1)],$$

where T is the suspended-sediment transport rate, in tons per day, and Q is the water discharge, in cubic feet per second. A higher order polynomial equation was not used because it could not be extrapolated to water-discharge values greater than the largest water discharge in the data set. A value of 1 was added to values of Q and T to eliminate the possibility of taking the logarithm of zero. The equations were determined by least-squares regression of logarithmically transformed values of mean daily water and suspended-sediment discharge, using Statistical Analysis System's¹ regression program (S.A.S. Institute Inc., 1982). Separate transport relationships were determined for each quarter for each stream. Analysis of covariance showed that separation of the suspended-sediment transport relationships by quarters was statistically valid. Both the slopes of the relationships and adjusted mean values of suspended-sediment discharge were significantly different at the 95-percent confidence level.

In all cases except the Upper Truckee River, second quarter, the slope of the line relating T to Q increased abruptly at a given discharge (different for each quarter for each stream). Therefore, once the value of Q at which the increase in slope took place was determined, separate relationships were determined for water discharges below and above that value. The water-discharge values at which the slope increase occurred generally were at or below base-flow levels for the study streams. Suspended-sediment transport by such small streamflows was negligible; nearly all significant suspended sediment was transported by larger flows. Therefore, even though small flows contributed little to the total suspended-sediment yield, it was necessary to separate them from larger flows in order to prevent the small discharge data from decreasing the slope of the more significant large discharge relationship (fig. 2). Walling (1977b) also recognized this problem in dealing with sediment-transport curves. Equations of the suspended-sediment transport curves for large discharges are shown in table 1.

Trout Creek during the fourth quarter of the water year was the only instance for which three distinct transport relationships were required -- one for small values of water discharge and two for moderate and large values of water discharge (table 1). The suspended-sediment-transport curves for moderate and large water discharges are nearly parallel but have different intercepts, indicating different levels of suspended-sediment transport under different

¹The use of the firm name in this report is for identification purposes only and does not constitute endorsement by the U.S. Geological Survey.

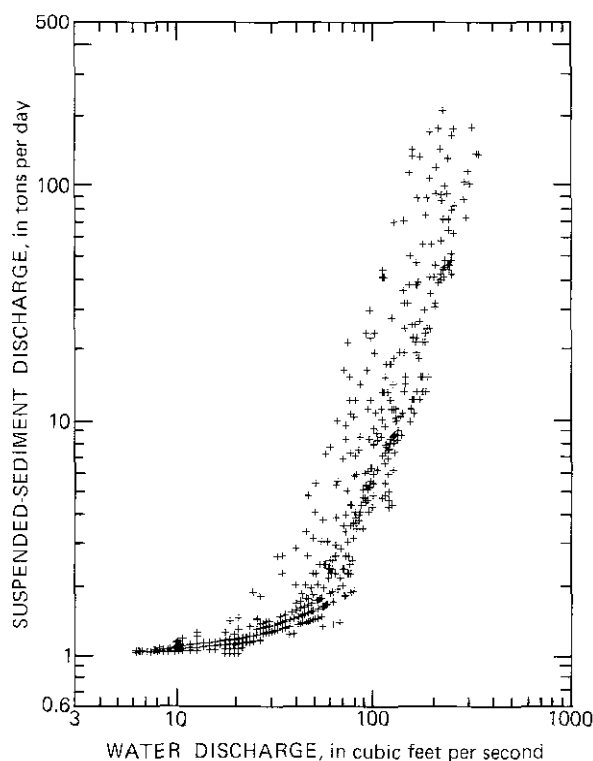


Figure 2. Plot of suspended-sediment discharge vs. water discharge for Blackwood Creek, third quarter (April-June). Separate transport relationships were determined for values of water discharge less than and greater than 44.0 cubic feet per second (table 1), in order to minimize the influence of small water-discharge and suspended-sediment-transport values on the relationship for large water discharges.

hydrologic conditions. Although the change in slope of the relationship from small to moderate discharges probably represents the change from virtually no transport to significant transport of suspended sediment, the abrupt increase in the level of the relationship at a water discharge value of 81.2 ft³/s may indicate a change in sediment availability reflected in the sediment-transport curve.

MAGNITUDE AND FREQUENCY OF SUSPENDED-SEDIMENT TRANSPORT

The contribution of each water discharge class to the total suspended-sediment load was obtained by combining the flow-duration distribution with the suspended-sediment-transport relationships, one discharge class at a time, according to the method described by Colby (1956, p. 24) and Miller (1951). The duration of flows that transported 10, 50, and 90 percent of the total suspended-sediment load were of particular interest because they enabled comparisons of the results of this study to those of other authors. The range of effective or "most-work" discharge, that discharge class which transported the largest quantity of suspended sediment during the period of study, also was determined for each stream.

Table 2 summarizes the magnitude and frequency of suspended-sediment transport for the study streams. Ten percent of the sediment was transported by flows greater than those which were equaled or exceeded from 0.062 to 0.72 percent of the time, or 0.22 to 2.6 days per year. Fifty percent of the sediment was transported by flows which were equaled or exceeded from 1.7 to 5.4 percent of the time (6.0 to 19.2 days per year), and 90 percent of the sediment was transported by flows greater than those which were equaled or exceeded 8.3 to 28 percent of the time (30 to 102 days per year). Third-quarter streamflow (April-June), due to snowmelt runoff, transported 73.8 percent of the total annual suspended-sediment load.

Estimated mean annual suspended-sediment load in the Lake Tahoe tributaries is summarized in table 3. Total suspended-sediment transport to Lake Tahoe by the study streams, which represent 36 percent of the basin area, during water years 1973 through 1981 was 44,000 tons.

DISCUSSION

The results of the magnitude-frequency analyses demonstrate that most suspended-sediment transport in the study streams was accomplished by flows of intermediate magnitude and frequency. Ninety percent of the suspended sediment was transported by flows which recurred more often than 1.4 days per year. Most suspended sediment was transported by frequently occurring streamflow. Although large volumes of material were transported by less frequent, large flows, the largest part of suspended sediment transported during the period of record was the result of flows within the normal yearly range of streamflow. In particular, flows associated with annual third-quarter snowmelt transported most of the suspended sediment.

Table 2. Magnitude and frequency of suspended-sediment transport for study streams.

Station	Flows below which 50% of sediment is transported (ft ³ /s)	Duration (percentage of time flow equaled or exceeded)	Recurrence interval (days/yr)	Flows below which 90% of sediment is transported (ft ³ /s)	Duration (percentage of time flow equaled or exceeded)	Recurrence interval (days/yr)
Upper Truckee River (10336610)	380	4.6	16.4	560	0.56	2.0
Blackwood Creek (10336660)	210	2.0	7.1	495	.062	0.22 (or 1 day/4.5 yr)
Ward Creek (10336676)	170	1.7	6.0	300	.17	0.61 (or 1 day/1.65 yr)
Trout Creek (10336780)	105	5.4	19.2	155	.72	2.6

Table 3. Average annual suspended-sediment yields
from study streams, water years 1973-81

Gaging station	Sediment yield [(ton/mi ²)/yr]	Drainage area (mi ²)
Upper Truckee River at South Lake Tahoe (10336610)	37.8	54.80
Blackwood Creek near Tahoe City (10336660)	129.9	11.16
Ward Creek at Highway 89 near Tahoe Pines (10336676)	58.4	9.70
Trout Creek near Tahoe Valley (10336780)	22.2	36.70

An understanding of suspended-sediment transport patterns is needed because transport of nutrients to Lake Tahoe, especially particulate phosphorus, has been shown to be associated with transport of suspended sediment (Goldman, 1974; Leonard and others, 1979). Leonard and others (1979) reported that patterns of particulate-phosphorus concentration in Ward Creek corresponded to seasonal water-discharge fluctuations. Peak particulate-phosphorus concentrations occurred during the period of snowmelt runoff. Ryden and others (1973) noted that particulate phosphorus adhered to clay-sized sediment particles. Because these particles are transported as part of the suspended-sediment load, the delivery of particulate phosphorus to Lake Tahoe is subject to the magnitude-frequency regime of suspended-sediment transport.

CONCLUSIONS

Most suspended sediment in the Lake Tahoe tributaries studied was transported by relatively frequent flows of moderate magnitude. The largest part of the suspended-sediment load was transported by flows associated with annual spring snowmelt. Data from this study indicate that, in a region where yearly snowmelt dominates the streamflow regime, sediment transport is less episodic than in those areas where rare storms have a greater impact.

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ISSUES IN SEDIMENT RESEARCH IN RIVERS OF THE U.S.

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ABSTRACT

Among the issues and problems that concern the sediment in rivers of the United States are: the effects of dams and reservoirs, the implications of sediment storage, and the importance of large infrequent storms. Dams and reservoirs have diminished by half the amount of sediment that the Mississippi River formerly transported to its delta. They have almost completely stopped the seaward transport of sediment by two other great rivers of the country, the Colorado and the Grande. The storage of sediment in river valleys at different time scales is a problem that requires urgent attention because, until we understand it more clearly, we are unable to predict the fate of many of the pollutant substances that are found in the nation's rivers. Furthermore, a better understanding of the frequencies and magnitudes of the very largest sediment-transporting events will improve our ability to predict the rates at which sediment is discharged by rivers.

INTRODUCTION

Sediment ranks high among the substances that are supplied to rivers by nonpoint sources. Part of the sediment supplied to rivers is a natural consequence of the steady geologic processes that erode the continents and transport the eroded material as sediment to the oceans. The rest of the sediment in rivers, which may be most of the sediment in some heavily affected rivers, is a consequence of the accelerated erosion that follows such human activities as forest clearing, crop farming, surface mining, and construction. Because most of the sediments supplied to rivers come from nonpoint sources, whether they are induced natural or artificially, the sources are difficult to identify, predict, and control.

Adding to the difficulties of predicting sediment inputs to rivers is the complexity of estimating the rate of sediment delivery. Although onsite erosion of specific types of soils under specific conditions of cultivation or other land uses can be predicted by using such tools as the Universal Soil Loss Equation (Wischmeier and Smith, 1965), predicting how much of the eroded soil will be delivered eventually to the channel of a neighboring stream still remains difficult. Sediment that has been eroded off upland fields often is deposited on hillslopes or in the upper parts of stream valleys before it reaches a water course. The length of time during which the sediment is stored in this manner can range from a few days to hundreds of years. Consequently, the sediment that one observes in a river channel today may represent episodes of erosion that took place several decades or even a century ago.

Once it reaches a stream channel, sediment may cause a number of problems. By raising the elevation of the channel bed, increased sedimentation can lead to increased flooding due to a decrease in carrying capacity of the stream channel. Sediment affects the maintenance of in-channel structures, navigation systems, and other works in the river environment. Furthermore, sediment particles adsorb many contaminants, such as pesticides, radionuclides, and

toxic metals that are transported, deposited, and stored as part of the sedimentary component of the riverine system.

Mean annual sediment loads discharged to the coastal zones reflect a number of artificial influences, not the least of which is the interruption of the downriver flow of sediment by dams and reservoirs. In spite of the large dams that have been built across its major tributaries, the Mississippi River still ranks sixth or seventh in the world in suspended sediment discharge to the oceans (Milliman and Meade, 1983, p. 2). Next in rank in the conterminous States is the Columbia River with an average load of 10 million tons per year before the 1980 eruption of Mount St. Helens, and an estimated load of 40 million tons per year after the eruption. Although over 140 million tons of suspended sediment from Mount St. Helens was discharged by the Cowlitz River into the Columbia River in the first 4 months after the eruption, this discharge has decreased considerably in the last few years. The additional sediment attributed to Mount St. Helens has declined to about 30 million tons per year.

Ten rivers of the United States that are important by virtue of their large sediment discharges or their large drainage areas are listed in Table 1. The present-day sediment discharges of three rivers that drain the glaciated peaks of the Alaska Range--the Copper, the Yukon, and the Susitna--rank respectively second, third, and fourth, among the rivers of the United States. The sediment discharges of the Mississippi River, the Rio Grande, and the Colorado River have been diminished by dams and reservoirs. Although it discharges three-quarters as much water to the ocean as the Mississippi River, the St. Lawrence River carries relatively little sediment because the Great Lakes act as natural sediment traps (Meade and Parker, 1985).

EFFECTS OF RESERVOIRS ON SEDIMENT LOADS

One of the most pervasive influences on sediment loads is exerted by the large number of dams and reservoirs that have been built across the rivers of the United States. Dams are built to impound water for various purposes, and the reservoirs they form interrupt the downriver flow of sediment. Although the river water that enters a reservoir is released eventually (through a power plant, into a diversion canal, or over a spillway), much of the sediment is trapped permanently in the reservoir. Nearly all reservoirs on major rivers of the United States trap at least one-half the river sediment that flows into them. Some of the largest reservoirs in the country, like Lake Powell and Lake Mead on the Colorado River, trap virtually all the sediment that flows into them.

While the effects of reservoirs on sediment loads are apparent in rivers in all parts of the country, they are most obvious in the large western rivers where the original sediment loads were naturally large and where the construction of dams has been especially intense. The effects of reservoirs on the quantities of the sediment loads transported by rivers are well described by extensive collections of data from three large western river systems--the Missouri-Mississippi, the Rio Grande, and the Colorado River--and a group of rivers in the Eastern United States (see Meade and Parker, 1985, for more detailed discussions and illustrations).

TABLE 1.--Discharge of suspended sediment to the coastal zone by
ten major rivers of the United States, about 1980
(ton/yr = tons per year)

	Mean annual sediment discharge (million ton/yr)
Rivers that discharge the largest sediment loads	
Mississippi-----	¹ 230
Copper-----	80
Yukon-----	65
Susitna-----	25
Eel-----	15
Brazos-----	11
Columbia	
before Mount St. Helens eruption-----	10
since Mount St. Helens eruption-(approximate)----	40
Rivers with large drainage areas:	
St. Lawrence-----	1.5
Rio Grande-----	0.8
Colorado-----	0.1

¹Includes Atchafalaya River.

STORAGE OF SEDIMENT IN RIVER SYSTEMS

The 1983 National Water Summary (U.S. Geological Survey Water-Supply Paper 2250, p. 68-69) emphasized the importance of sediment storage in the overall picture of erosion and sedimentation. It points out that, on a national scale, the amount of sediment delivered to the oceans by rivers was only about 10 percent of the total amount eroded off the uplands of the country and that 90 percent of the soil eroded in the country was being stored somewhere between erosion sites and the sea. A fraction of this 90 percent is being stored in reservoirs but most of it is stored in other places such as on hillslopes, flood plains, and other parts of stream valleys. The implications of the large amount of sediment storage are enormous; for example, because many of the toxic materials that travel in streams, such as metals, radionuclides, pesticides, and other organic substances are adsorbed tightly onto sediment particles, any accurate prediction of the fate of toxic substances in a stream will require an understanding of what is happening to the sediment. Sediment storage is difficult to predict, however, because it involves many different sedimentary processes, operating at a variety of different time scales (Walling, 1983).

Short-term (seasonal) storage of sediment in river channels is probably easier to understand and predict than long-term storage in river systems. Some of the short-term changes in storage of suspended sediment can be seen in the lower reaches of the Mississippi River in Louisiana. No dams obstruct these

reaches of the Mississippi, no tributaries bring in sediment, and no outlets drain sediment away until it reaches the mouth of the river. Any downriver changes that are observed in the discharges of suspended sediment represent deposition of material onto the riverbed or resuspension of material from the riverbed. The short-term pattern shows sediment being deposited and stored on the riverbed at lower flows and being resuspended and flushed out to the Gulf of Mexico on higher flows. Analogous patterns of seasonal storage and remobilization of sediment have been observed and described from rivers that range in size from small (Emmett and others, 1983; Meade, 1985) to the largest in the world (Meade and others, 1983, p. 1139-1140; 1985).

Questions involving the seasonal storage and resuspension of sediment in the lower Mississippi River that need to be studied include the following: What is the long-term balance between storage and resuspension--in the long run, is more sediment being stored than resuspended, or vice versa?; how does the seasonal storage affect the pollutants that are adsorbed on the sediment particles?

A well-known case of long-term movement and storage of sediment in a river system is that of the hydraulic-mining debris in the Sacramento River valley of California (Gilbert, 1917; Kelley, 1959). Between 1855 and 1885, enormous quantities of sediment were washed into some of the tributaries of the Sacramento River during hydraulic mining for gold. The resulting problems that developed downstream (flooding, filling of navigation channels, destruction of flood-plain farms) became so serious that hydraulic mining was curtailed by a court decision in 1884. By that time, however, a large mass of sediment, characterized as a "wave" by G.K. Gilbert (1917), was already into the stream channels and was moving slowly down the tributaries and into the Sacramento River. As the mass of sediment advanced, it raised the levels of the channel beds, much as an ocean wave raises the level of the sea as it passes through. Bed levels rose 19 feet in the tributary Yuba River at Marysville and nearly 11 feet in the Sacramento River at the town of Sacramento. The riverbeds at these towns reached their greatest elevations 10 to 20 years after the mining was stopped, and then they declined steadily to their previous elevations during the next 30 to 40 years. All in all, the great wave of hydraulic-mining debris took nearly a century to pass through the channels of the Sacramento River system and come to rest finally in San Francisco Bay.

This pattern, however, applied only to the sediment in and near the river channels. It did not apply to the debris that had overflowed onto the flood plains. The hydraulic-mining debris that was carried out of the river channel during floods and deposited on the flood plains was sufficient in many places to cover entire houses and orchards (Kelley, 1959, p. 134-135, 203-204). Most of this debris still remains today where it was deposited a century ago. The time required to remove sediment from storage on the flood plain is much longer than the century that was required to remove the debris from the main river channels. Flood-plain deposits are removed mainly by erosion of channel banks as streams slowly migrate laterally, a process that proceeds at a substantially slower pace than the vertical removal of material stored in the bottom of the river channel.

Many of the problems associated with the prediction of long-term sediment storage are demonstrated in a study carried out on Coon Creek, a small stream that drains 140 square miles of southwestern Wisconsin (Trimble and Lund, 1982). Originally covered by forests, Coon Creek basin was settled by European immigrants and cleared for farming about 1850. As the forests were cleared and the land was plowed, a cycle of erosion and sedimentation began, the consequences of which are still strongly in effect today. In 1933, after about 80 years of land-management practices that resulted in excessive erosion, soil-conservation efforts were begun in earnest. These efforts still continue.

Details of the Coon Creek study can be found in two recent publications (Trimble and Lund, 1982; Trimble, 1983). The study demonstrates the complexity of the storage problem. The time scales of storage here are so long and the storage sites so diverse that it is difficult to even begin to construct mathematical models to predict the eventual rates of sediment movement. As outlined in a recent summary by Walling (1983), the problems of sediment delivery and long-term storage in river valleys are among the principal challenges for future studies of sediment.

EFFECTS OF INFREQUENT LARGE STORMS ON SEDIMENT TRANSPORT

In many rivers of the conterminous United States, a large proportion of the sediment load is transported in only a small proportion of the time; for example, within any individual year, more than one-half of the sediment load for the year is likely to be transported in only 5 or 10 days. Also, over a period of many years, a large proportion of the long-term sediment load may be transported in response to a few large, but infrequent, storms.

The frequencies of suspended-sediment discharge within individual years and the importance of infrequent large storms in producing large sediment load can be demonstrated by the daily suspended-sediment discharge records for three stations - Eel River at Scotia, Calif., Delaware River at Trenton, N.J., and Juniata River at Newport, Pa. These three data sets were selected because each contained the effects of a large storm (whose recurrence interval was longer than the period of sediment record), and each contained sufficient data from years of more average sediment discharge to place the effects of the storm into some kind of comparative context. Among the three rivers, nearly half of a year's sediment usually is discharged in one percent of the time, and nearly 90 percent usually is discharged in 10 percent of the time.

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NOTE: Page 4-168 is the last printed page in Volume 1 of the Proceedings for the 4th Federal Interagency Sedimentation Conference.