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STEWART L. UDALL, *Secretary*

GEOLOGICAL SURVEY

Thomas B. Nolan, *Director*

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Fluvial Sediments—a Summary of Source, Transportation Deposition, and Measurement of Sediment Discharge

By BRUCE R. COLBY

CONTRIBUTIONS TO GENERAL GEOLOGY

GEOLOGICAL SURVEY BULLETIN 1181-A

General description of fluvial sediments from source and entrainment through the processes of transportation and deposition and brief notes on sediment measurements

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ILLUSTRATIONS

TABLES

SYMBOLS

A	Constant in the Meyer-Peter and Müller bedload equation.
A_1	Constant in the Bagnold bedload equation.
a	Distance above streambed, usually used by Einstein as equal to $2D$, the thickness of the bedlayer.
B	Constant in the Meyer-Peter and Müller bedload equation.
B_1	Parameter used in the Bagnold bedload equation; varies with characteristics of the sediment particles.
C	Chezy C.
c_a	Sediment concentration, by weight, at a distance a above the streambed.
c_y	Sediment concentration, by weight, at a distance y above the streambed.
D	Particle size expressed as a representative diameter.
D_m	Particle size used in Meyer-Peter and Müller bedload equation and for uniform sediment equal to D.
D_{90}	Particle size at which 90 percent of the particles by weight are smaller.
d	Depth of flow.
E	Average annual sheet erosion or soil loss.
F	Basic soil-erodibility factor.
g	Gravity constant.
i_b	Fraction, by weight, of bed material in a size range.
K	Constant whose different uses are indicated by numerical subscripts.

K_r Measure of resistance to flow equal to $\bar{u}/(R^{2/3}S^{1/2})$.
K_r' Comparable measure of flow resistance due to particle resistance only.
k Universal turbulence constant under the assumptions for which z was derived.
L Length of land slope in direction of flow.
P_{30} Maximum precipitation during a 30-minute period, in inches.
q_B Bedload discharge, in dry weight of sediment per unit width.
q_B' Bedload discharge per unit width, by weight underwater.
R Hydraulic radius.
R' Hydraulic radius with respect to the sediment particles; that is, the hydraulic radius that could be used to compute velocity if all the resistance to flow were due to the sediment particles and none to the configuration of bed and banks.
R_e Relative erosion rate for different types of vegetal cover and agricultural practice.
S Energy gradient for which surface slope is frequently substituted.
S_L Slope of the land surface.
$\bar{u}$ Mean velocity.
v_s Fall velocity of sediment particles.
y Distance above the streambed.
z Abbreviation for $v_s/(k\sqrt{gRS})$, an exponential measure of the vertical distribution of sediment of uniform fall velocity.
β Declination of the bed surface below horizontal.
γ_f Specific weight of the fluid.
γ_s Specific weight of the solid sediment particles.
Θ_F Measure of the apparent tangential stress due to action of external forces on the fluid.
Θ_t Value, defined by extrapolation, of Θ_F at which sediment particles will begin to move over a rippled bed.
Φ Intensity of bedload transport. An asterisk as a subscript indicates that the function is for one size fraction of bed material.
Ψ Function for correlating effect of flow with intensity of sediment transport. An asterisk as a subscript indicates that the function is for one size fraction of bed material.

DEFINITIONS

Ability of a stream to transport sediment. Power of a stream to discharge sediment at a cross section must be expressed in terms of a particle-size distribution as well as a rate of sediment discharge.

Aeolian sediment. Sediment that is transported by, or suspended in, air or that has been deposited in beds by air currents.

Antidunes. Symmetrical waves of the bed sediment of a stream exist below symmetrical surface waves. Surface waves and bed waves may be relatively stable over a period of several minutes, or they may build up, break, and re-form rapidly.

Backwater curve. Surface profile of a stream whose mean velocity is gradually decreasing with downstream distance.

Bedlayer. A thin layer through which the bedload is discharged; commonly assumed to be only a few grain diameters thick.

Bedload. Sediment that is transported in a stream by rolling, sliding, or skipping along the bed and very close to it; that is, within the bed layer.

Bed-material load. Sediment that is transported by a stream and that consists of particle sizes large enough to be found in appreciable quantities at the surface of the streambed.

Discharge of water or sediment. Time rate of movement of volume or weight of the water or sediment past a point or through a cross section.

Dunes. Irregularly spaced low mounds of loose sand which travel slowly downstream as the result of sand being moved along their comparatively gentle upstream slopes and being deposited on their steeper downstream slopes.

Fall velocity of a sediment particle. Rate of settling of the particle in a fluid.

Fine-material load. Usually called wash load; consists of sediment so fine that it is about uniformly distributed in the vertical and is only an inappreciable fraction of the sediment on the streambed. Its upper size limit at a particular time and cross section is a function of the flow as well as of the sediment particles.

Fluvial sediment. Sediment that is transported by, or suspended in, water or that has been deposited in beds by water.

Plane bed. A reasonably smooth bed of unconsolidated sediment that is practically free from ripples, dunes, antidunes, and bars. Ideally, the resistance to flow over a plane bed is due to the size and roughness of the individual sediment particles and not to irregularity of the bed configuration.

Median diameter or median particle size. Midpoint in the size distribution of a sediment for which half the particles by weight are larger and half are smaller.

Recurrence interval of a hydrologic event of a particular size. Average time between events that equal or exceed that size.

Sediment. Fragmental material that originates from the disintegration of rocks and is transported by, suspended in, or deposited by water or air or is accumulated in beds by other natural agencies.

Sediment concentration. Ratio of dry weight of sediment to the total weight of water-sediment mixture.

Sheet erosion. Erosion that removes a roughly equal depth of sediment from each small unit area of land surface.

Trap efficiency of a reservoir. Proportion, usually expressed as a percentage, of the sediment inflow that is retained by the reservoir.

Velocity-weighted concentration for a vertical or cross section. Concentration that is obtained from samples whose rate of collection at all sampled points was about in proportion to velocity. Such a concentration can be used with rate of water discharge to compute sediment discharge for the sampled part of a vertical or cross section.

CONTRIBUTIONS TO GENERAL GEOLOGY

FLUVIAL SEDIMENTS—A SUMMARY OF SOURCE, TRANSPORTATION, DEPOSITION, AND MEASUREMENT OF SEDIMENT DISCHARGE

By Bruce R. Colby

ABSTRACT

This paper presents a broad but undetailed picture of fluvial sediments in streams, reservoirs, and lakes and includes a discussion of the processes involved in the movement of sediment by flowing water.

Sediment is fragmental material that originates from the chemical or physical disentegration of rocks. The disintegration products may have many different shapes and may range in size from large boulders to colloidal particles. In general, they retain about the same mineral composition as the parent rocks.

Rock fragments become fluvial sediment when they are entrained in a stream of water. The entrainment may occur as sheet erosion from land surfaces, particularly for the fine particles, or as channel erosion after the surface runoff has accumulated in streams.

Fluvial sediments move in streams as bedload (particles moving within a few particle diameters of the streambed) or as suspended sediment in the turbulent flow. The discharge of bedload varies with several factors, which may include particle size and a type of effective shear on the surface of the streambed. The discharge of suspended sediment depends partly on concentration of moving sediment near the streambed and hence on discharge of bedload. However, the concentration of fine sediment near the streambed varies widely, even for equal flows, and, therefore, the discharge of fine sediment normally cannot be computed theoretically. The discharge of suspended sediment also depends on velocity, turbulence, depth of flow, and fall velocity of the particles.

In general, the coarse sediment transported by a stream moves intermittently and is discharged at a rate that depends on properties of the flow and of the sediment. If an ample supply of coarse sediment is avaliable at the surface of the streambed, the discharge of the coarse sediment, such as sand, can be roughly computed from properties of the available sediment and of the flow. On the other hand, much of the fine sediment in a stream usually moves nearly continuously at about the velocity of the flow, and even low flows can transport large amounts of fine sediment. Hence, the discharge of fine sediments, being largely dependent on the availability of fine sediment upstream rather than on the properties of the sediment and of the flow at a cross section, can seldom be computed from properties, other than concentrations based directly on samples, that can be observed at the cross section.

Sediment particles continually change their positions in the flow; some fall to the streambed, and others are removed from the bed. Sediment deposits form locally or over large areas if the volume rate at which particles settle to the bed exceeds the volume rate at which particles are removed from the bed. In general, large particles are deposited more readily than small particles, whether the point of deposition is behind a rock, on a flood plain, within a stream channel, or at the entrance to a reservoir, a lake, or the ocean.

Most samplers used for sediment observations collect a water-sediment mixture from the water surface to within a few tenths of a foot of the streambed. They thus sample most of the suspended sediment, especially if the flow is deep or if the sediment is mostly fine; but they exclude the bedload and some of the suspended sediment in a layer near the streambed where the suspended-sediment concentrations are highest.

Measured sediment discharges are usually based on concentrations that are averages of several individual sediment samples for a cross section. If enough average concentrations for a cross section have been determined, the measured sediment discharge can be computed by interpolating sediment concentrations between sampling times. If only occasional samples were collected, an average relation between sediment discharge and flow can be used with a flow-duration curve to compute roughly the average or the total sediment discharges for any periods of time for which the flow-duration curve and the sediment-discharge relation can be assumed to apply.

Although many agencies have collected sediment records, the available information is inadequate to determine sediment concentrations or discharges in more than a rough way for many streams or parts of streams in the United States.

INTRODUCTION

Water in lakes, reservoirs, and streams may have a content of sediment ranging from only a trace to more than 500,000 ppm (parts per million), by weight, at peak concentrations in a few streams. Unfortunately, the physical laws of sediment transportation are incompletely known even for the narrow range of particle size commonly used in a laboratory flume. They are especially inexact and inadequately defined for the complex problem of transportation of the wide range of sediment sizes that are carried by natural streams. A generalized idea of the present knowledge and currently accepted theories may, however, help toward an understanding of the movement of sediment particles in water.

The purpose of this paper is to present a generalized concept of fluvial sediments and their distribution and movements from the points of entrainment to the mouths of rivers. Theories of sediment transportation are stressed, particularly with reference to erosion and deposition of the sediments, but no phase of sediment transportation is discussed in detail. The nature and availability of sediment records in the United States are indicated briefly.

The paper begins with a short explanation of the sources and entrainment of fluvial sediments and then presents, in order, the theories of sediment transportation, the deposition of sediment from

flow, methods of measuring sediment discharge, and the computation and the general availability of sediment records in the United States.

SOURCE AND ENTRAINMENT

Sediment may be defined as fragmental material that originates from the disintegration of rocks and is transported by, suspended in, or deposited by water or air or is accumulated in beds by other natural agencies. Some sediment results from volcanic activity, and some from the breaking down of rocks by earthquakes. However, the principal source of rock fragments that become fluvial sediments is the disintegration of rocks of the earth's crust. The rock fragments at or near the land surface form soils after weathering and mixing with organic materials. They become fluvial sediment when they are entrained in water, as when unconsolidated particles are eroded by sheet flow or by channel flow. Some aeolian sediment falls on water surfaces as loose particles or in raindrops and thus becomes fluvial sediment. Source and entrainment of fluvial sediments will, therefore, be discussed briefly in terms of the disintegration products of rocks, erosion of unconsolidated sediments by water, and derivation of fluvial sediments from aeolian sediments.

DISINTEGRATION OF ROCKS

Rocks at or near the surface of the earth are continually subjected to chemical and physical actions that may decompose or change them and that are likely to break the rocks down to small fragments. According to Twenhofel (1939, p. 5), "* * * the most important chemical processes are oxidation, carbonation, and hydration." Mechanical forces such as crustal movements, grinding, impact, root action, temperature changes of the rocks, and thawing and freezing of water are effective in breaking down rocks to fragments. Rocks and rock particles may be abraded by moving water and by sediment in air or water. The chemical and physical forces work slowly, but huge quantities of fragmental material are gradually derived from rocks. Much of this material, especially that in the smallest size ranges, is susceptible to transportation by currents of water or air.

The rock fragments range in size from large boulders to colloidal particles and may be cohesionless or cemented together either loosely or tightly. They may be well rounded by abrasion or have the sharp edges that characterize freshly broken chips of the parent rock. Usually the mineral composition of the fragments is about that of the rocks from which the fragments came.

Size-gradation scales are needed to place nomenclature and terminology on a uniform basis, and many size classificaions have been proposed. The designation commonly used consists of classes and

subclasses in which the upper size limit of each subclass is two times the lower limit. Table 1 gives the size limits for each class and subclass of a scale recommended by the American Geophysical Union Subcommittee on Sediment Terminology (Lane, 1947, p. 937).

The size, shape, and mineral composition of the rock particles that become fluvial sediments may all be significant factors in some practical stream problems, such as the sorption of radioactivity on the particles. According to Sayre, Guy, and Chamberlain (1963), sorption is roughly proportional to the surface area of sediment particles. If spherical shape and comparable packing are assumed for sediments of different sizes, the surface area of equal weights of sediment particles varies inversely with diameter, whence 1-micron clay particles would have 1,000 times the surface area of an equal weight of 1-mm sand grains. Generally, clay particles tend to be

TABLE 1.—*Scale of sizes, in metric and English units*

Class and subclass	Metric		English
	Millimeters	Microns	Inches
Boulders			
Very large boulders	4, 096–2, 048		160–80
Large boulders	2, 048–1, 024		80–40
Medium boulders	1, 024– 512		40–20
Small boulders	512– 256		20–10
Cobbles			
Large cobbles	256–128		10–5
Small cobbles	128– 64		5–2. 5
Gravel			
Very coarse gravel	64–32		2. 5–1. 3
Coarse gravel	32–16		1. 3– . 6
Medium gravel	16– 8		. 6– . 3
Fine gravel	8– 4		. 3– . 16
Very fine gravel	4– 2		. 16– . 078
Sand			
Very coarse sand	2. 000–1. 000	2, 000–1, 000	. 078– . 039
Coarse sand	1. 000– . 500	1, 000– 500	. 039– . 020
Medium sand	. 500– . 250	500– 250	. 020– . 0098
Fine sand	. 250– . 125	250– 125	. 0098– . 0049
Very fine sand	. 125– . 062	125– 62	. 0049– . 0024
Silt			
Coarse silt	. 062– . 031	62–31	. 0024– . 0012
Medium silt	. 031– . 016	31–16	. 0012– . 00061
Fine silt	. 016– . 008	16– 8	. 00061– . 00030
Very fine silt	. 008– . 004	8– 4	. 00030– . 00015
Clay			
Coarse clay size	. 004 – . 0020	4 –2	. 00015– . 000076
Medium clay size	. 0020– . 0010	2 –1	. 000076– . 000038
Fine clay size	. 0010– . 0005	1 –0. 5	. 000038– . 000019
Very fine clay size	. 0005– . 00024	. 5– . 24	. 000019– . 000010

flatter and may have different packing than sand grains, but the dominant effect of size on surface area is obvious. Sorption by clays may be due mainly to ion exchange, whereas physical sorption rather than chemical sorption is usual for coarser sediments. Furthermore, Sayre, Guy, and Chamberlain (1963) state,

> The chemical and mineralogical nature of sediments, particularly of clays, is also important in determining the extent of sorption. The cation-exchange capacity of the clay, and the position of the adsorbed cations in the lyotropic series with respect to that of the radioactive cations in solution would be significant factors in this regard.

EROSION OF UNCONSOLIDATED SEDIMENTS BY WATER

Erosion by water may be divided into sheet erosion and channel erosion; however, like many distinctions that are made in sediment nomenclature, no sharp division exists between the two. In a sense, true sheet erosion, which calls for removal of a roughly equal depth of sediment from each small unit area of a plot or field, will seldom occur, because runoff tends to flow over the land surface in many small rills rather than in sheets of flow. However, a practical distinction is made here on the basis of the way in which rates of the two types of erosion are frequently computed. Rate of sheet erosion or soil loss is usually approximated from characteristics of the soil, climate, vegetal cover, and topography of a field or small area. Rates of channel erosion, including gully erosion, are usually estimated directly or indirectly from the characteristics of the flow and channel rather than from factors that relate to a relatively unbroken land surface.

SHEET EROSION

Some of the sediment, especially the fine sediment, carried by streams has recently been removed from land surfaces by sheet erosion. The many factors that affect the rate of sheet erosion have been studied extensively by soil scientists and agriculturists because of the damage caused by such erosion. Empirical relations have been developed for estimating average rates of soil loss through long periods of time, but the quantity of sheet erosion over a watershed of appreciable size during individual storms generally cannot be computed satisfactorily.

The physical characteristics of a soil—including particle size, cohesiveness, porosity, and moisture content—partly determine the rate of sheet erosion. The permeability of the soil and subsoil affects the proportion of precipitation that remains to flow over the ground and thus possibly to cause erosion. The amount of loose dust at the soil surface largely determines the concentration of fine sediment in the first runoff during a storm period. Some of these soil characteristics vary widely from storm to storm, but on the average they are

relatively stable over a period of years. Usually, computations of soil loss are based on the determined erodibility of a soil for certain standard conditions of land slope, precipitation, and vegetal cover such as those given by Musgrave (1947, table 3, col. 7). Adjustments for departures from the standard conditions are then applied to this basic erosion rate for a soil.

Vegetation and vegetal litter intercept some rain and may greatly reduce the force with which the raindrops strike the soil; they also help to hold the soil in place and to reduce the velocity of water flowing over the ground. Deep roots increase the permeability of the subsoil, so that more water drains away below the surface. Generally, vegetal cover reduces the soil moisture in the root zone at the beginning of storms during the growing season, so that more of the initial precipitation can be retained in the soil. An idea of the large effect of vegetal cover is obtained from Musgrave (1947, p. 135–136), who gives a relative erosion rate of 100 percent for continuous row crops, 15 to 40 percent for small grains, and less than 1 percent for hay, pasture, woodland, and forest.

Probably, the steeper a land slope, the larger the fraction of precipitation that will run off over the ground if all other pertinent factors remain constant. Steep slopes also tend to cause high velocities of flow over the soil surface. The farther the water flows over the soil surface, the higher its velocity normally becomes, because of its increased depth, the more and its greater opportunity it has for picking up sediment. Hence, the degree of land slope and, to a smaller extent, the length of the slope have significant effects on average rates of sheet erosion.

Precipitation on a soil surface tends to loosen soil particles and make them readily available for transportation. Therefore, the energy of the rainfall has a direct effect on rate of erosion. Rain that falls in large drops is generally more effective in causing erosion than rain that falls as fine drops. Hail is particularly effective in loosening soil particles. Thus, an intense rain composed of large drops, and perhaps driven by a high wind, not only tends to loosen soil particles but also usually supplies adequate flow to carry them away.

Relations given by Musgrave (1947) for computing rates of sheet erosion from small areas such as plots or small fields can be expressed in a general equation of the form

$$E=FR_e\left(\frac{P_{30}}{1.25}\right)^{1.75}\left(\frac{S_L}{10}\right)^{1.35}\left(\frac{L}{72}\right)^{0.35} \qquad \text{(a)}$$

in which

E is the long-time average annual soil loss or rate of sheet erosion, in tons per acre or in inches of average depletion;
F is the basic soil erodibility factor, in the same units as E;
R_e is a relative erosion rate for different types of vegetal cover and agricultural practice;
P_{30} is the maximum precipitation, in inches, in a 30-minute period for a recurrence interval of 2 years;
S_L is the average slope of the land surface, in percent; and
L is the length of the slope, in feet, in the direction of flow.

This general type of equation has been used in many forms. It requires modification for different areas and is, of course, not precise but does indicate the significance of principal factors that determine average rates of sheet erosion over a long period of time. It is not generally applicable for computing quantities of erosion for a particular year or for individual storms, because the physical characteristics of the soil, the soil moisture, and the vegetal cover—especially for cultivated crops—all vary widely from storm to storm and from year to year.

Sheet erosion generally removes the dust and fine soil from the ground surface. Factors such as the readily available supply of fine sediments on the surface, vegetal cover, and steepness of land slopes largely determine the concentration of the fine fluvial sediment. Vegetal cover, degree of land slope, permeability of the soil, and rates and total quantities of precipitation are dominant factors that affect the volume of runoff and total quantity of the fine sediment that will be transported from a drainage area by storm runoff.

CHANNEL EROSION

Runoff from a land surface quickly concentrates into small streams, which increase in size as they join together. The flow of these streams tends to erode the channel banks and beds wherever sediment is available of particle sizes that are not already being carried to the full ability of the streams to transport them. Channel erosion, which is the entrainment of sediment from the bed or banks of a stream, may be general or local.

General channel erosion occurs when the flow of a stream has the ability to transport much more of the available sediment than it is already carrying. Thus, channel erosion is common below reservoirs until the released water has had time to adjust its carrying ability to the available sediment. The adjustment may be through degradation of the channel to bank and bed sediments that, because of size or cohesiveness, cannot be transported in significant quantities by the

flow. It may be through selective sorting of these sediments or through reduction of the stream slope as a result of degradation near the reservoir. Generally, the adjustment is made through a combination of a decrease of ability to transport sediment and a change in the effective availability of the sediment. Most natural streams are relatively well adjusted to the sediment loads that they carry, but the release of abnormal quantities of flow into a channel that has much fine sediment readily available in bed and banks may result in spectacular erosion. For example, Schroeder and Miller (written communication, 1953) of the U.S. Bureau of Reclamation, in a paper entitled "A Plan of Channel Erosion Control, Fivemile Creek, Riverton Project, Wyoming," estimated that 27,000 acre-feet of sediment was removed from about 30 miles of Fivemile Creek during the first 15 years of use as a wasteway for the Wyoming Canal. (See figs. 1, 2.)

Most stream channels erode locally even though their flows are in general equilibrium with the available sediment loads. A winding stream tends to scour the outside of bends; but much of the eroded sediment, especially the larger particle sizes, may be quickly deposited on the inside of the bends. A stream locally constricted in width tends to erode sediment at the narrow section during periods of high flow but may quickly deposit the sediment where the channel again widens. Either singly or together, the abrasive effect of flow and the softening effect of saturation may cause streambanks to slough into the channel and thus become subject to rapid removal by the flow.

Gullies are a conspicuous form of channel erosion. They develop readily, especially on steep slopes where an ephemeral stream flows over unconsolidated sediments of sand sizes or smaller. Rates of

FIGURE 1.—Fivemile Creek downstream from the crossing of Wyoming Canal before waste water caused accelerated erosion. Photograph by U.S. Bureau of Reclamation.

FIGURE 2.—Fivemile Creek downstream from the gaging station near Shoshoni, October 1947. The extreme width of the channel and the high vertical banks resulted from accelerated erosion.

gully erosion are extremely variable and are not readily correlated quantitatively with physical characteristics of a watershed. They vary with particle size and cohesiveness of the sediments in the banks and bed of the gully, with vegetal cover, with quantity of runoff, and with the concentration of the runoff in both time and space. Most of the runoff that erodes a particular gully may originate outside the gully. A computation of nearly 2 inches of vertical depth of erosion per year has been made (Woodburn, 1949, p. 22) for a gully whose bed and banks were mostly sand. The average runoff of streams near this gully is about 20 inches annually, of which about 15 inches comes from direct surface runoff. Many gullies are in areas where average annual runoff is less than 1 inch.

If the bed slope flattens where the stream leaves a gully, much of the coarse sediment in transportation may be deposited and the stream may spread laterally until most of its velocity and sediment-carrying ability are lost. Hence, much of the sand and soil aggregates eroded from gullies moves only a short distance, unless a reasonably well-defined and continuous channel of only gradually decreasing slope extends away from the gully.

The particle sizes of sediment available and the ability of streams to transport the particles determine the size composition of the fluvial sediments that are eroded from gullies or from other stream channels Much fine sediment transported during periods of low flow may be derived directly from channel erosion.

FLUVIAL SEDIMENTS ENTRAINED FROM AEOLIAN SEDIMENTS

At times, much sediment is carried by wind, especially in arid and semiarid regions. Chepil and Woodruff (1957, p. 110) computed quantities of sediment transported during some dust storms in Kansas and eastern Colorado during 1954 and 1955. For a time on March 10, 1954, when visibility was 0.05 mile at Syracuse, Kans., they estimated sediment transportation of nearly 48,000 tons per hour per mile perpendicular to the direction of the wind. Other computed rates of transportation of aeolian sediment were, of course, much lower.

Aeolian sediments readily become fluvial sediments when they fall on water surfaces or where sand dunes migrate into bodies of surface water, but usually only small amounts of fluvial sediment in a stream have come directly from the air. However, streams frequently pick up sediment where they flow across aeolian deposits. As much aeolian sediment consists of very fine particles, the fluvial sediments derived from aeolian deposits may be more significant in some hydraulic and water-use problems than their usually small quantities would indicate.

TRANSPORTATION

A generalized theory of sediment transportation is explained in the following few paragraphs as a background for more detailed and quantitative discussions of the movement of sediment in flowing water. The explanation is both qualitative and relative and appilies to relations within a reach of channel rather than to relations between streams.

Some sediment that reaches a flowing stream is transported by the water, and some may be deposited along the channel; other sediment is eroded from the channel. Fine particles are transported mainly or entirely in suspension through the supporting action of the turbulence of the water and may move without appreciable deposition. Coarse particles also may travel in suspension, may be rolled or skipped along the streambed, or may be transported alternately by the two methods. The finest sediments move with about the velocity of the flowing water. They may pass directly with the water from place of erosion to points far downstream. Much of this reasoning follows that of Einstein, Anderson, and Johnson (1940). Large sediment particles are likely to be deposited temporarily or semipermanently at places along the stream. At any time and place on the streambed, the probability of deposit and the probable length of time before moving again are largely functions of particle size. Much of the coarsest sediment may be at rest far more of the time than it is moving, even though it is at the surface of the streambed and exposed to the flow.

If appreciable amounts of coarse sediments are deposited along the channel, they are likely to be rather uniformly available for pickup by subsequent flows.

In general, the concentrations of both fine and coarse suspended sediments within a reach of stream channel increases as flow increases. However, the concentration increases for fine sediment are much less consistent than those for coarse sediment. The concentration of the fine sediment usually increases as flow increases because most high flows result from rainfall or snowmelt that erodes fine sediment from the land sufaces; but the peak concentrations of fine sediment may not coincide with the peaks of flow, and the largest runoffs do not necessarily produce the highest concentrations of fine sediment. The concentration of the coarse sediments increases as discharge increases, principally because velocities tend to be faster and flow more turbulent at high discharges.

Another way of thinking of sediment transportation within a particular reach of a stream is that the discharge of fine particles is controlled by the available supply of such particles, and the supply is usually much less than the stream can transport. The supply of the coarse particles is usually greater than the stream can transport, and the discharge of these particles is regulated mainly by the ability of the stream to transport them. Thus, the concentration of the coarse sediments at a section is a function of factors such as velocity and water temperature, which can be measured at the section. In contrast, the concentration of the fine particles is relatively independent of the flow characteristics at a section, because even comparatively low flows are easily able to transport the available fine sediments.

The study of sediment transportation by streamflow is, therefore, complex, because many variables are involved and some of them are difficult to express mathematically except, perhaps, by curves. The variables relate not only to available supply of the sediment but also to sizes, shapes, and densities of the particles; velocities of flow; channel widths, depths, and slopes; bank roughness and bed configuration; and density, temperature, and at times even chemical composition of the water. An average particle size or mean velocity may be an inadequate measure, respectively, of particle sizes of a sediment or of velocity at a cross section, because the distribution about the average has significant effects. Most factors affecting sediment discharge change not only with time and with distance along a channel but also with depth and with lateral distance at an individual cross section. Consequently, at a cross section, each sediment size (more precisely, sediment of each particular fall velocity) has its own lateral and vertical distribution, which usually differs significantly from the distribution of sediment of other sizes. The first major problem in a

study of sediment transportation is, then, to reduce it to terms simple enough to be explained and understood. This simplification usually requires (a) classification of the sediment loads according to size or mode of transportation and (b) selection of a relatively few factors that have major effects and cannot be disregarded. However, the major factors in one type of sediment study may differ appreciable from those for another type.

In the following explanation of sediment transportation, the total sediment load of a stream is divided into parts that are somewhat differently related to characteristics of channel and flow.

The term "load" in this paper designates sediment that is carried by a stream, sometimes by a particular mode of transportation as, for example, bedload. "Load" is a less specific term than "discharge," which here means a time rate of movement of dry weight of sediment through a cross section.

On the basis of mode of transportation, the total sediment load of a stream can be divided into two parts. One part is the bedload, which consists of the sediment that moves by skipping, sliding, or rolling and that always remains very close (generally within a few grain diameters for uniform sediment) to the streambed. The other part is the suspended-sediment load, which may be anywhere within the turbulent flow and is maintained in the flow by the upward components of turbulent currents or by colloidal suspension if the sediment particles are very small.

The total sediment load of a stream can be divided into two other parts on the basis of general relation to the flow. One part is the fine-material load, sometimes called wash load, which consists of particles so fine that they are not found in appreciable quantity in the streambed. The other part is the bed-material load, which is composed of particle sizes found in appreciable quantity at the surface of the streambed. The bed-material load includes nearly all the bedload. It may be transported anywhere within the depth of flow. Except at low velocities relative to the particle sizes or at very shallow depths of flow, more bed material is discharged as suspended sediment than as bedload.

The dividing size between fine-material load and bed-material load is determined by the size distribution of the sediment at the surface of the streambed. As the size distributions of both the sediment load of a stream and the bed sediment usually are continuous, the dividing size is partly arbitrary. For many sand-bed streams, a dividing size of 0.062 mm can be used; however, the dividing size varies with velocity, water temperature, and the quantities and sizes of sediment available upstream. The dividing size will usually be different for flow in the main channel than for overbank flow. Al-

though uncertainties exist, the distinction is important because different factors are significant in the computation of the discharge of fine-material load and of bed-material load (Einstein and others, 1940).

If the flow is at equilibrium with its sediment load, the presence of sediment of a given size in appreciable quantity at the surface of the bed indicates that more sediment of that size is available than the stream will transport. The lack of an appreciable quantity of fine sediment at the surface of the bed indicates that the fine sediment is being carried through the section without significant deposition. However, some of the fine sediment may deposit near the banks or in other places where the velocity is abnormally low. Also, some fine sediment may be trapped among the coarser particles below or at the surface of the streambed.

BEDLOAD

A study of sediment transportation logically begins with theories of bedload movement, because the discharge of bedload provides a basis for the computation of bed-material discharge and is also closely related to processes of erosion and deposition.

Theoretically, bedload discharge should correlate more simply with characteristics of the flow and the bed sediment than should either discharge of suspended bed material or discharge of fine material. All theories of sediment transportation, however, require experimental verification, and bedload discharge is hard to measure. Hence, the development of relations for computing bedload discharge has been slow. Three currently used relations will be discussed in terms of their broad meanings.

Meyer-Peter and Müller (1948) developed an equation for bedload discharge, which becomes, for channels having negligible bank resistance,

$$\gamma_f (K_r/K_r')^{3/2} RS = A(\gamma_s - \gamma_f) D_m + B(\gamma_f/g)^{1/3} (q_B')^{2/3} \tag{b}$$

in which

γ_f and γ_s are the specific weights, respectively, of the fluid and the sediment particles;

K_r is a measure of resistance to flow and equals the mean velocity $\bar{u}$ divided by $(R^{2/3}S^{1/2})$;

K_r' is a comparable measure of the flow resistance due to particle resistance only;

R is the hydraulic radius, which equals depth of flow when the bank resistance is negligible;

S is the energy gradient;

A and B are constants that may be used as 0.047 and 0.25, respectively;
D_m is the particle size of uniform sediment and equals ΣDi_b for a bed sediment of different sizes (i_b is the fraction, by weight, of sediment of a size D);
g is the gravity constant; and
q_B' is the bedload discharge per unit width, by weight under water.

K_r' for metric units can be computed from $26/(D_{90})^{1/6}$. D_{90} is the particle size, in meters, for which 90 percent of the bed sediment by weight is finer.

In foot-pound-second units and for γ_s equal to 165 pounds per cubic foot, equation (b) can be written—

$$(K_r/K_r')^{3/2}RS = 0.077D_m + 0.0050(q_B')^{2/3}$$

For convenience, the bedload discharge per foot of width can be expressed in terms of the dry weight per foot of width q_B through the relation—

$$(q_B')^{2/3} = [(165-62.4)/165]^{2/3}(q_B)^{2/3}$$

and equation (b) becomes—

$$(K_r/K_r')^{3/2}RS = 0.077D_m + 0.0036(q_B)^{2/3} \qquad \text{(c)}$$

According to equation (c), no bedload moves until the left-hand side exceeds $0.077D_m$. For high velocities and small particle sizes, $0.077D_m$ becomes relatively small. If it is disregarded and (K_r/K_r') is constant, the computed bedload discharge varies as $(RS)^{3/2}$ or as the third power of the mean velocity if the Chezy C is constant.

This equation can be understood better by applying it to the computation of bedload discharge for a particular cross section of a stream. The bed sediment at the selected section has a median diameter of 0.40 mm, D_m in equation (c) is 0.45 mm, and D_{90} is 0.66 mm. Bedload discharges were computed for five different velocities and depths of flow; the three lowest velocities were over dune beds, and the two highest over plane beds. The computed bedload discharges increase rapidly and consistently as the mean velocity increases, but a break exists in their relations to RS. (See fig. 3.) Evidently, RS is a poor measure of the discharge of bedload unless it is adjusted by some factor, such as $(K_r/K_r')^{3/2}$, that corrects for the effect of changes in resistance to flow owing to changes in channel roughness.

Except for low velocities in relation to depth of flow, the curve of computed bedload discharges on figure 3 is at least roughly applicable for many streams that have beds of sand.

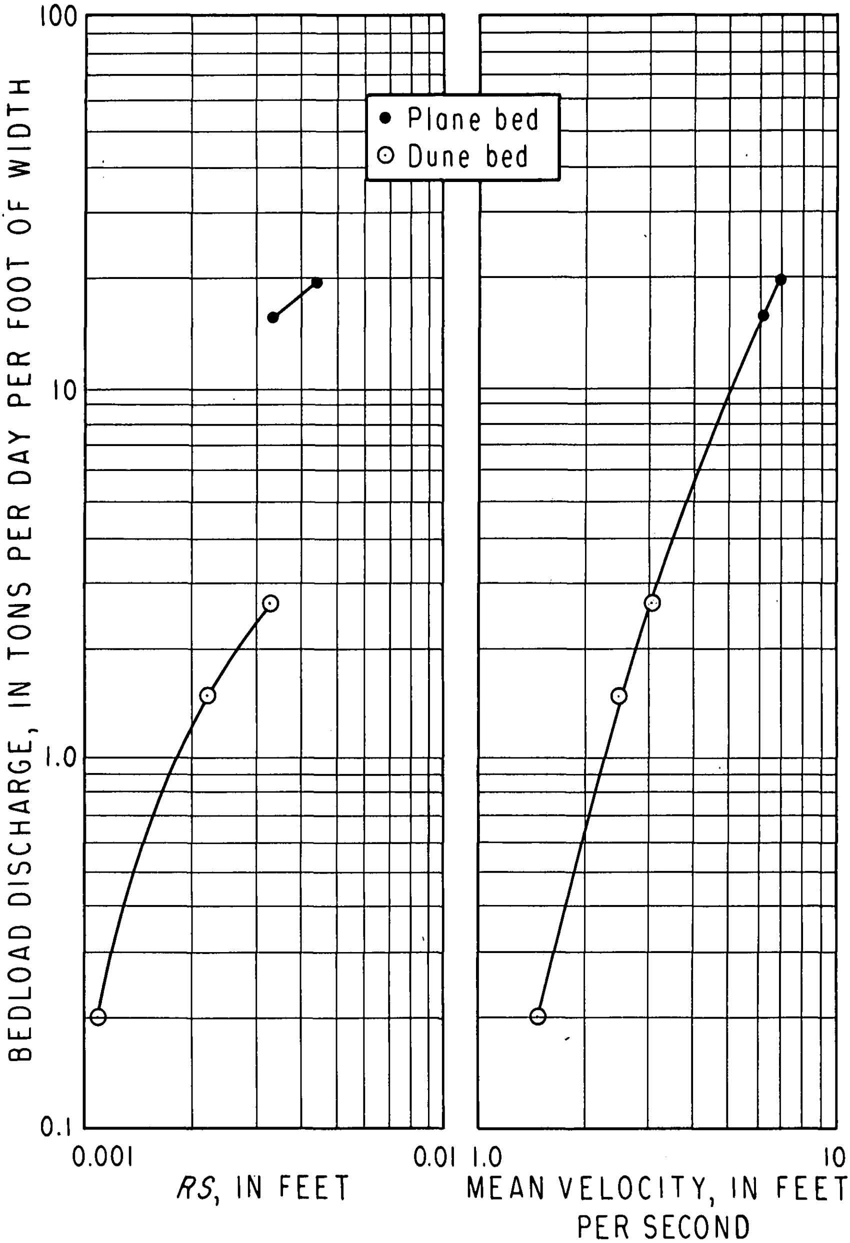

FIGURE 3.—Bedload discharges, computed with the Meyer-Peter and Müller equation, as functions of *RS* and mean velocity.

Einstein (1950) suggested a comprehensive procedure for the computation of bed-material discharge from characteristics of the flow and of the bed sediment. In this procedure, the part of the bed-material discharge that is transported as bedload was computed on the basis of the probability of movement of particles at the bed surface as related to intensity of flow. Einstein's bedload equations for uniform sediment can be written

$$\Phi=\frac{q_B}{\gamma_s}\left(\frac{\gamma_f}{\gamma_s-\gamma_f}\frac{1}{gD^3}\right)^{1/2}$$

$$\Psi=\frac{\gamma_s-\gamma_f}{\gamma_f}\frac{D}{R'S}$$

$$\Phi=f(\Psi)$$

For constant specific weights and one consistent system of units of measurement

$$q_B=K_1D^{3/2}\Phi$$

$$=K_1D^{3/2}f\left(\frac{K_2D}{R'S}\right)$$

and K_1 and K_2 are constants. R' is the hydraulic radius that could be used to compute velocity if all the resistance to flow were due to the sediment particles and none to the configuration of the bed and banks.

Einstein used R' in equations based on work by Keulegan (1938), but R' can equally well be used in a Manning velocity equation. That is, R' can be stated in terms of R, the actual hydraulic radius (or average depth of flow for negligible bank friction), as

$$\bar{u}=K_rR^{2/3}S^{1/2}$$

$$=K_r'(R')^{2/3}S^{1/2}$$

in which

$\bar{u}$ is the mean velocity.

Whence

$$R'=R(K_r/K_r')^{3/2}$$

Einstein's equation for q_B then may be written, with K_3 a constant, as

$$q_B=K_3D^{3/2}f\left[\frac{D}{(K_r/K_r')^{3/2}RS}\right]$$

Thus, Einstein's parameter for the effect of flow on bedload discharge of uniform particle size is the same as that used in the Meyer-Peter and Müller equation.

Einstein stressed the need for the adjustment from R to R', an adjustment that ties the bedload discharge more closely to mean velocity than to RS. However, he computed the relation between R and R' through the use of a curve that may be a good average but sometimes does not apply well.

If the bed sediment is not uniform in size, the Einstein procedure requires considerable adjustment of Φ and Ψ. The adjusted quantities for individual size classes or size fractions of bed sediment are Φ_* and Ψ_*. Einstein's graph of Φ_* as a function Ψ_* (fig. 4) shows a rapid decrease of bedload discharge when the velocity, as measured by $R'S$, becomes low in relation to the particle size; that is, when Ψ_* becomes large.

From a different approach than that used by Einstein, Bagnold (1956) derived a relation for bedload discharge considerably like Einstein's. For bed sediment of uniform size, his equation (Bagnold, 1956, eq. 42c) is

$$q_B = A_1 B_1 K_1 D^{3/2} \sqrt{\cos\beta}(\text{Ⓗ}_F - \text{Ⓗ}_t)\text{Ⓗ}_F^{1/2} \qquad \text{(d)}$$

in which

A_1 is a constant that may be about 9;
B_1 is a constant that varies slowly with particle size;
K_1 is the same constant that was used previously in $q_B = K_1 D^{3/2}\Phi$;
Ⓗ_F is a measure of the apparent tangential stress due to action of external forces on the fluid and equals $\frac{\gamma_f}{(\gamma_s - \gamma_f)}\frac{RS}{D}$;
Ⓗ_t is that value of Ⓗ_F (determined by extrapolation) at which sediment particles begin to move on a rippled sand bed;
β is declination of the bed surface below the horizontal, and $\cos\beta$ generally can be considered to equal unity and be disregarded.

Except for the use of R instead of R', Bagnold's Ⓗ_F is the reciprocal of Einstein's Ψ. Also, for a particular grain size, constant densities, and moderately high velocities so that Ⓗ_t becomes relatively insignificant, the bedload discharge according to equation (d) is about proportional to $(RS)^{3/2}$. For these stated conditions, the bedload discharge is also proportional to the third power of the mean velocity if the Chezy C is constant.

A major difference between Bagnold's equation for computing bedload discharges and equations of Einstein or Meyer-Peter and Müller is that Bagnold makes no provision for the effect of changing

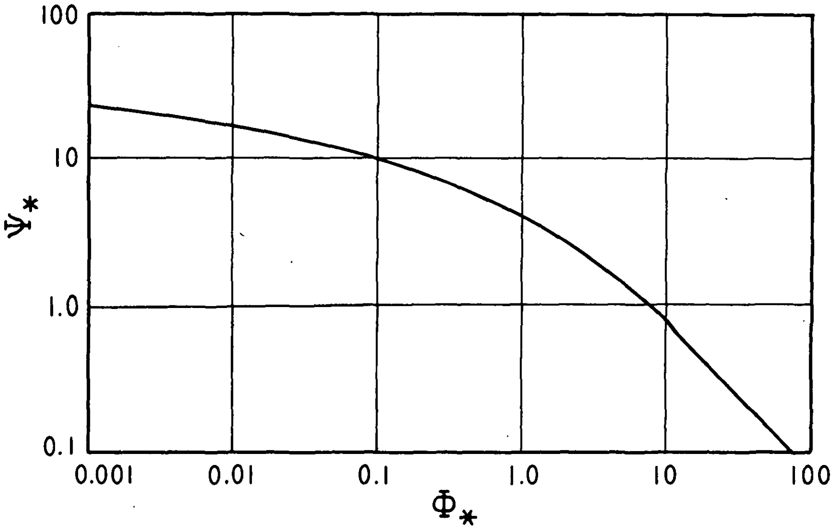

FIGURE 4.—Relation of Ψ_* and Φ_* for individual size fractions. From Einstein (1950, fig. 11).

resistance because of changing bed configuration. That is, he uses neither an R' instead of R nor a K_r/K_r' ratio. If RS is as inadequate a measure of bedload discharge as figure 3 indicates, Bagnold's equation may not be as generally applicable as the Einstein equation or the Meyer-Peter and Müller equation.

BED-MATERIAL LOAD

Mathematical theories of sediment transportation are somewhat unsatisfactory for computing bed-material discharge and are generally impractical for computing fine-material discharge. The theories are significant, nevertheless, partly because they provide some insight into the discharge of both fine and coarse sediment and partly because they apply, at least qualitatively, to the deposition of both fine and coarse sediment.

Most bed material is discharged in suspension rather than as bedload. The bedload moves only in a very thin layer near the bed and is transported by the comparatively slow velocities that exist near the bed, whereas the bed-material load may be distributed throughout the depth of flow.

The concentration of suspended bed material at the top of the bed layer (the layer through which bedload is discharged) probably is continuous with its concentration in the bed layer. In other words, the concentration of moving sediment in the bed layer probably provides a base to which the concentration of suspended bed material adjusts itself. Einstein (1950, p. 38–39) used this assumed relation in a method for computing the discharge of suspended bed material.

Einstein (1950) computed the discharge of suspended bed material of a given size fraction (size range) by integrating the product of point velocity and concentration of the size fraction from the top of the bed layer to the water surface. He gave values of the integrals (Einstein, 1950, figs. 1, 2) to be used. Vertical distribution of point velocity was computed from equations stated by Keulegan (1938). The vertical distribution of concentration of sediment of a narrow size range was based on an equation that was given by Rouse (1937, p. 535) and used by Einstein (1950) in the form

$$c_y/c_a=\left(\frac{d-y}{y}\,\frac{a}{d-a}\right)^{v_s/(k\sqrt{gRS})}$$

$$=\left(\frac{d-y}{y}\,\frac{a}{d-a}\right)^{z} \qquad \text{(e)}$$

in which

- c_y and c_a are concentrations, by weight, of suspended sediment of a narrow size range at distances y and a above the bed;
- d is the depth of flow and equals the hydraulic radius R for two-dimensional flow;
- z is an abbreviation for $v_s/(k\sqrt{gRS})$;
- v_s is the fall velocity of sediment of a representative size for the narrow size range; and
- k is the universal turbulence constant under the assumptions for which equation (e) was derived.

Einstein equated c_a, the concentration at the top of the bed layer of thickness a, to the average concentration in the bed layer. (See p. A16–A17 for a brief discussion of the method used by Einstein for computing bedload discharge and, hence, concentration within the bed layer.)

The total bed-material discharge of a size fraction is the sum of the bedload discharge and the suspended bed-material discharge for that size fraction; and the total bed-material discharge is, of course, the sum of the discharges of all the size fractions of bed material. Einstein's procedure for computing total bed-material discharge is applicable only for long reaches of fairly uniform channel for which representative cross sections and accurate energy gradients can be determined. The streambed of the reaches should be unconsolidated sediments that generally shift readily, at least at moderate to high flows. Computed vertical distributions of velocity and concentration are likely to be somewhat inaccurate over relatively plane beds and are highly questionable over beds of sand dunes.

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The whole Einstein procedure is inexact and laborious. However, it probably gives the most detailed and comprehensive picture that is available of the significant factors in bed-material discharge.

Anyone who makes many computations by the Einstein procedure soon learns that small errors in determining mean velocity usually cause large errors in computed bed-material discharges, because an increase in velocity is generally associated with a large increase in computed bed-material discharge. Therefore, both the energy gradient and the roughness of the channel need to be correctly determined. As Brooks (1958, p. 588–591) has shown, the average relationship used by Einstein (1950, fig. 5) to adjust for roughness due to bed configuration is sometimes inexact.

The problem of resistance to flow is both significant and complex. When water flows over a bed of cohesionless sand, the flow not only moves sediment particles but also molds the configuration of the sand bed. Vanoni and Brooks (1957) have shown that major changes in resistance to flow are due to changes in bed configuration. The Chezy *C* may be about double for a plane bed what it is for a dune bed.

If velocity over a sand bed is increased and if depth, particle size, and water temperature are kept constant, an initially plane bed without sediment movement may shape itself successively into ripples, dunes, a plane bed, and antidunes. Some intermediate bed configurations also exist at least during periods of changing flow. Even at steady flow, the streambed at a particular time may have different configurations at different places. Also, bank roughness may markedly affect the flow when the width-to-depth ratio is comparatively small or when the banks are relatively rough as compared to the streambed. Because of the difficulty of evaluating roughness, bed-material discharges can be more simply and probably more accurately related to mean velocity than to depth and energy gradient.

The major effect of mean velocity on bed-material discharges can readily be shown for Niobrara River near Cody, Nebr. (See fig. 5.) The bed-material discharges were determined at a constricted cross section where practically all the bed material is in suspension and can be measured. Widths and mean velocities were measured at a section about 1,900 feet upstream from the constricted section. No adjustment was applied for changes in water temperature, although temperature has an appreciable effect.

Many factors such as particle size and size distribution, depth, water temperature, lateral and vertical distribution of velocity, concentration of fine sediment, and channel roughness affect the relation between bed-material discharge and mean velocity. However, the relation for the Niobrara River gives a good rough idea of the bed-

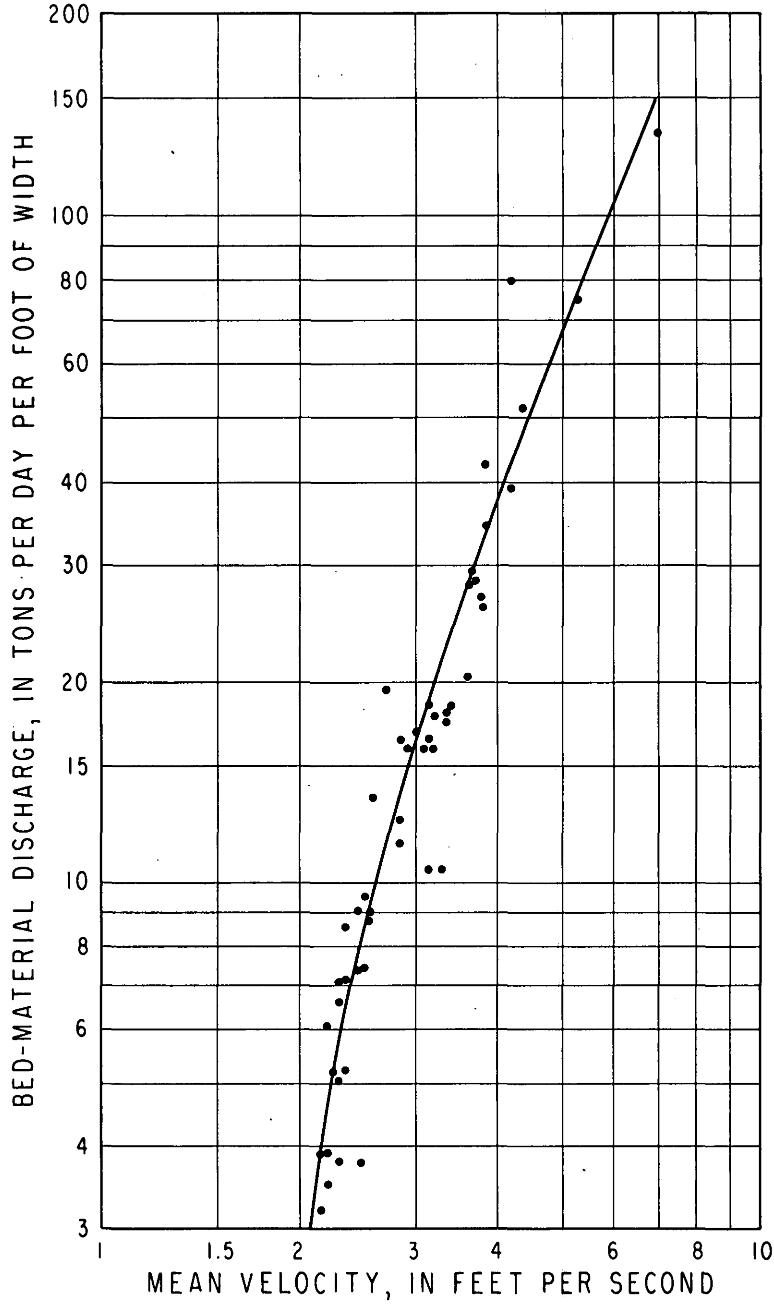

FIGURE 5.—Relation of bed-material discharge to mean velocity for Niobrara River near Cody, Nebr.

material discharge per foot of width of many natural streams that have beds of sand. It will not usually apply in laboratory flumes because of a depth effect, which becomes increasingly pronounced at low velocities. However, it does apply to channel reaches in which the energy gradient may be changing so rapidly that an average energy gradient would be unrepresentative.

FINE-MATERIAL LOAD

The procedure developed by Einstein cannot be used to compute discharge of fine material. So little fine sediment is at the surface of the bed that its percentage usually cannot be determined accurately by sampling, and the fine particles also are to some extent shielded by the larger particles. A bedload discharge of these fine particles is generally so small that it is disregarded, and its determination is much to inaccurate to be used as a basis for computing total discharge of fine particles through the cross section. Thus, although the same basic principles of sediment transportation apply to both fine-material discharge and bed-material discharge, practical difficulties prevent the use of these principles in the actual computation of fine-material discharge.

The part of the fine-material discharge that is eroded from the stream channel should correlate with stream stage or discharge. However, for many streams, most fine sediment usually comes from erosion of land surfaces rather than from channel erosion.

In general, the fine sediment is eroded from the land surface by overland flow during storms. The sheet erosion during individual storms is difficult to compute because of changes from time to time in the condition of the soil and because the erosive effect of precipitation during a particular storm is hard to evaluate. Too, some fine soil that is eroded may be deposited without traveling very far, especially if the rills overflow or are discontinuous. Fine-material discharges can, therefore, seldom be computed satisfactorily from rates of sheet erosion during individual storms.

Experience and general principles of sediment erosion and transportation indicate a few inexact relations between the fine-material discharge and the streamflow. Broadly, the concentration of fine material is high during direct storm runoff. Therefore, the concentration of fine sediment of a stream such as Powder River at Arvada, Wyo., can be as high at the occasional low flows that result from direct runoff as at high flows, although the average concentration of fine sediment is much lower at low flow than at high flow (fig. 6). If the distance of travel from the point of erosion is short or the stream channels contain little flow prior to the storm runoff, the peak concentration of fine material usually coincides with the peak flow or

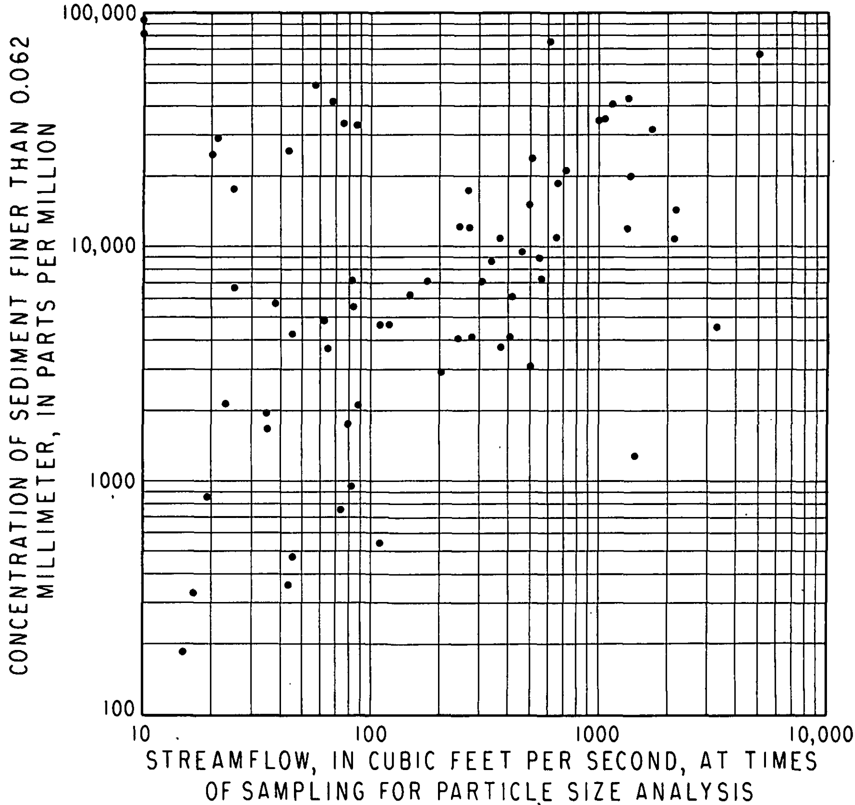

FIGURE 6.—Concentration of fine sediment plotted against flow of Powder River at Arvada, Wyo. At low flow, samples are more likely to be collected for size analysis when the concentration is high than when it is low. Hence, the observed concentrations at low flow may not be representative.

somewhat precedes it. Peak concentration of fine material early in the runoff is consistent with the idea that loose soil particles at the beginning of a storm will be eroded by the first direct runoff of appreciable amount. However, the flow from one tributary of a stream or from one part of a drainage area may be markedly lower or higher in concentration than the rest of the flow, and the time of arrival of such unrepresentative flow may determine the peak of fine-material concentration. The peak of the concentration of fine material may even lag far behind the peak of the flow (Heidel, 1956) if the fine material originated far upstream and if just before the storm runoff the stream channel contained large volumes of water having low sediment concentration.

In most streams, the concentration of fine material for a particular rate of flow varies with time of the year (Colby, 1956, figs. 6, 13, 24,

28, 31), because storm intensities, soil conditions, vegetal cover, and snowmelt vary seasonally. The amount of seasonal variation thus depends mainly on climate and on agricultural practices.

Current methods for computing the fine-material discharge give only an approximate idea of the long-time average discharge of fine sediment and of the variation of fine-material discharge with precipitation or runoff. As most sediment transported by streams is fine material, theories of sediment transportation are helpful but very inadequate for computing total sediment discharge, which should, therefore, usually be based principally on sediment-discharge measurements.

DISTRIBUTION OF SEDIMENT AT STREAM CROSS SECTIONS

The relative distribution of sediment of each particle size at a vertical depends on a parameter that is theoretically equal to $v_s/(k\sqrt{gRS})$ for flow over a plane bed. The fall velocity v_s is a function of size, shape, and roughness of the particles, water temperature, and the relative densities of the particles and the water. Because it varies widely for the usual particle sizes suspended in most streams, the vertical distributions of the individual particle sizes differ greatly. (See fig. 7.)

An increase in water temperature causes an increase in the fall velocities of sediment particles (fig. 8) and has an effect similar to an increase in particle size. However, the percentage increase in fall velocity as the temperature rises is small for particles larger than 1 mm, so that vertical distribution of the larger particles is little affected by temperature changes. Except at low stream velocities, a temperature increase also has comparatively little effect on the vertical distribution of the fine sediment because the fall velocities are always small whatever the temperature may be. The effect of changes in water temperature on fall velocity, vertical distribution, and discharge of sediment is large (Straub, 1954) for sediment in the size range from 0.1 to 0.4 mm.

Flocculation of the fine particles in suspension may considerably increase the effective fall velocities; but this effect, like the temperature effect for fine sediments, usually is highly significant only at low stream velocities.

High concentration of clay increases the effective viscosity of the water-sediment mixture and thus decreases the fall velocity. The effect is comparable to decreasing the water temperature or to decreasing the particle sizes of the fluvial sediments. However, high concentrations of bentonite, for example, may decrease the effective viscosity much more than any usual temperature change in natural streams decreases the viscosity.

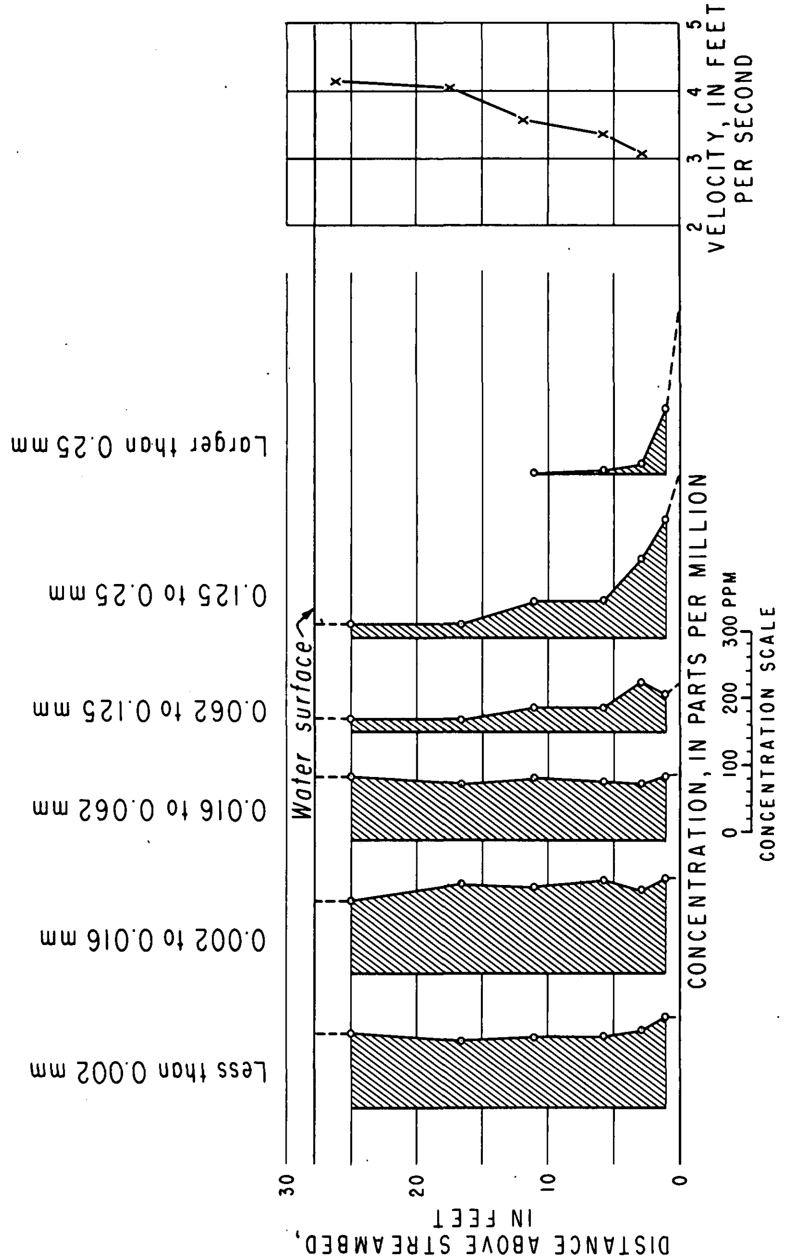

FIGURE 7.—Distribution of different sediment sizes at a vertical in Mississippi River at St. Louis, Mo., April 24, 1956.

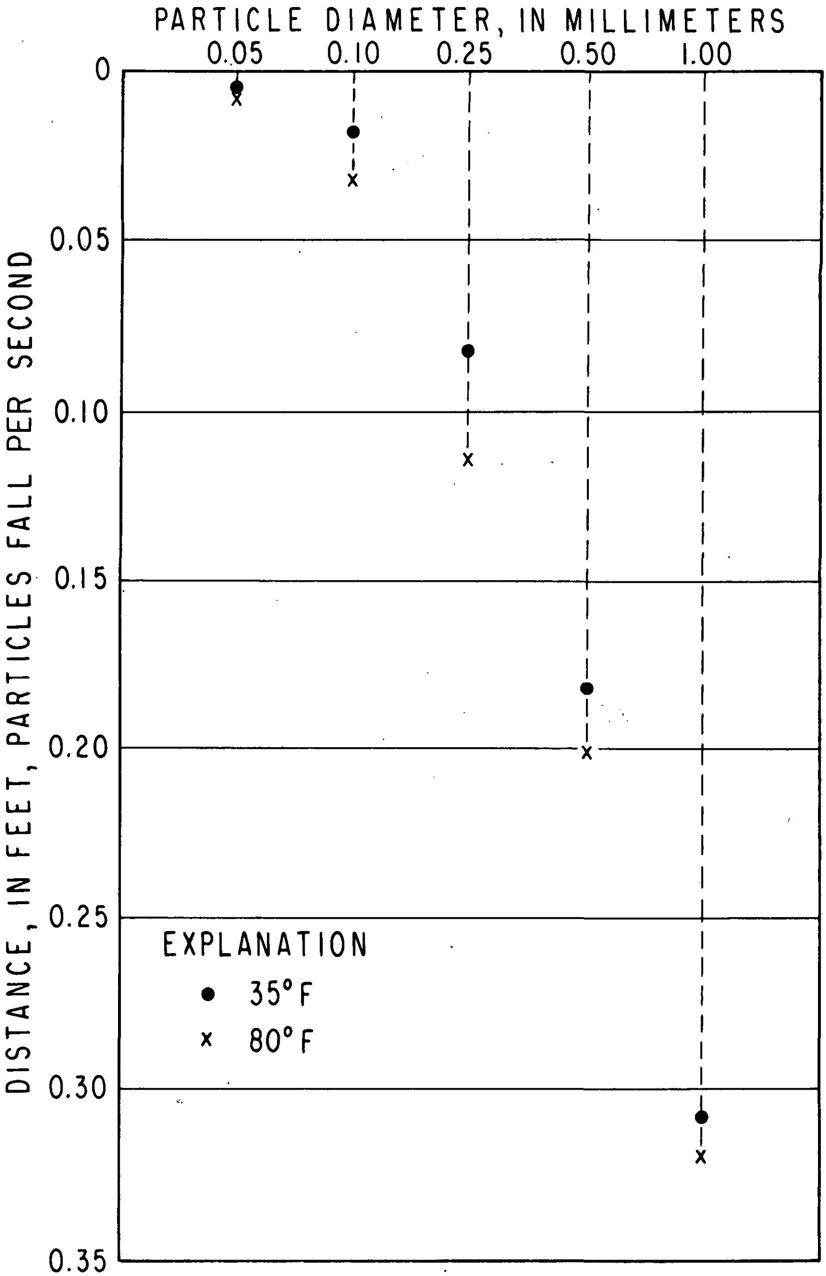

FIGURE 8.—Approximate effect of temperature on the fall velocities of quartz particles based on Hubbell and Matejka (1959, fig. 38).

The quantity $k\sqrt{gRS}$ increases in general as velocity increases. Therefore, an increase in velocity has much the same effect on vertical distribution of sediment as has a decrease in particle size. For example, on July 16, 1951, and on July 9, 1956, the reported temperature of Mississippi River at St. Louis, Mo., differed only 1°; but the vertical distribution of the sediment coarser than 0.062 mm was considerably more uniform on July 16, 1951, when the mean velocity was about 8.0 feet per second than on July 9, 1956, when the mean velocity was about 3.5 feet per second. (See fig. 9.)

Lateral distribution of sediment at cross sections follows less clear-cut rules than vertical distribution. If two streams join, their waters may not mix completely for a long distance downstream. The fine sediments remain with their associated flows and become well mixed only when the waters also become well mixed. Although some large sediment particles may migrate somewhat with respect to the particles of water that surround them, the lateral migration is probably a minor factor in the mixing of sediments downstream from the junction of two streams.

In turbulent streams whose waters are thoroughly mixed, the fine sediment is rather uniformly distributed laterally unless the flow is rapidly increasing, and then the water near the banks may have relatively low concentrations. To some extent and for equal depths, the coarse sediment is more concentrated where the velocities are highest and is almost always less concentrated where velocities are low. In particular, slow flows over an appreciable part of the stream width or shallow overbank flows transport little coarse sediment.

The lateral distribution of different size fractions of sediment in Niobrara River near Valentine, Nebr. (fig. 10), illustrates some of these general principles, but curvature of the stream and upstream depths and velocities affect the distribution of coarse sediment at this section as well as at many others in natural streams.

RATE OF TRAVEL OF SEDIMENT PARTICLES

The fine sediment travels with about the velocity of the water. Hence, methods for computing the dispersal of fine sediment may be useful in predicting the dispersal of many kinds of wastes in a stream. That is, the time of travel of clay and silt from the source of contamination in a stream to a critical point downstream should be practically the same as the time of travel of the dissolved minerals or the particular elements of water. Also, the time distribution of sediment concentration at a downstream section during a short period of fine-sediment loading should be much the same as the time distribution of contamination that might be associated with the water. Some fine sediment may be deposited in slack water at least semiperma-

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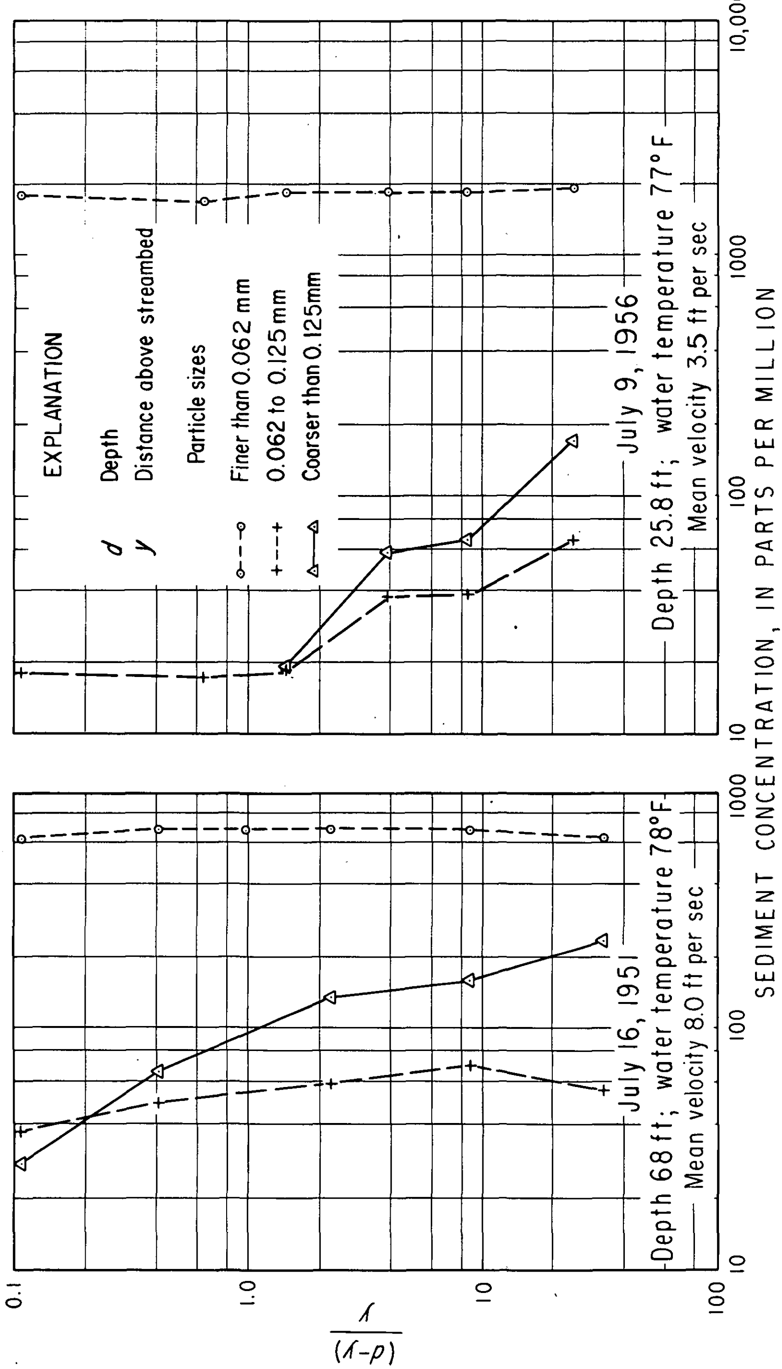

FIGURE 9.—Effect of velocity on the vertical distribution of different sizes of suspended sediment at a vertical of Mississippi River at St. Louis, Mo.

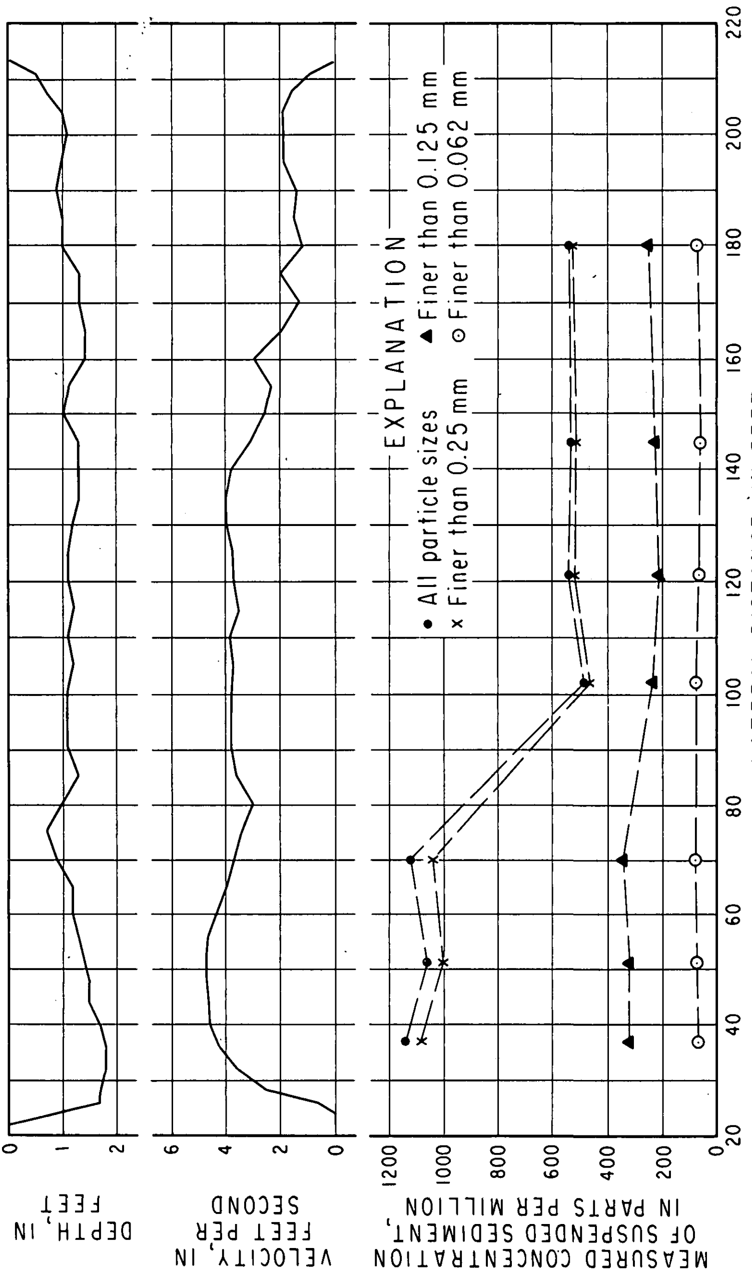

FIGURE 10.—Lateral distribution of measured sediment concentration of Niobrara River near Valentine, Nebr.

nently, and some may be trapped in the streambed. Also, the relatively small amount of wastes caught by the coarse sediment may move slowly downstream. Little is known about the rate of movement of individual particles of bed material, although these particles generally do not move continuoulsy but spend intermittent periods at rest.

Some flood peaks on the Bighorn River in Wyoming and Montana have been shown by Heidel (1956) to travel about 1.6 to 1.7 times as fast on the average as the peak concentrations of sediment, which was mostly fine particles. He concluded that the fine sediment traveled at about the speed of the water.

The water that formed the peak flows (fig. 11) at Kane, Wyo., and Bighorn, Mont., 55 and 170 miles, respectively, downstream from

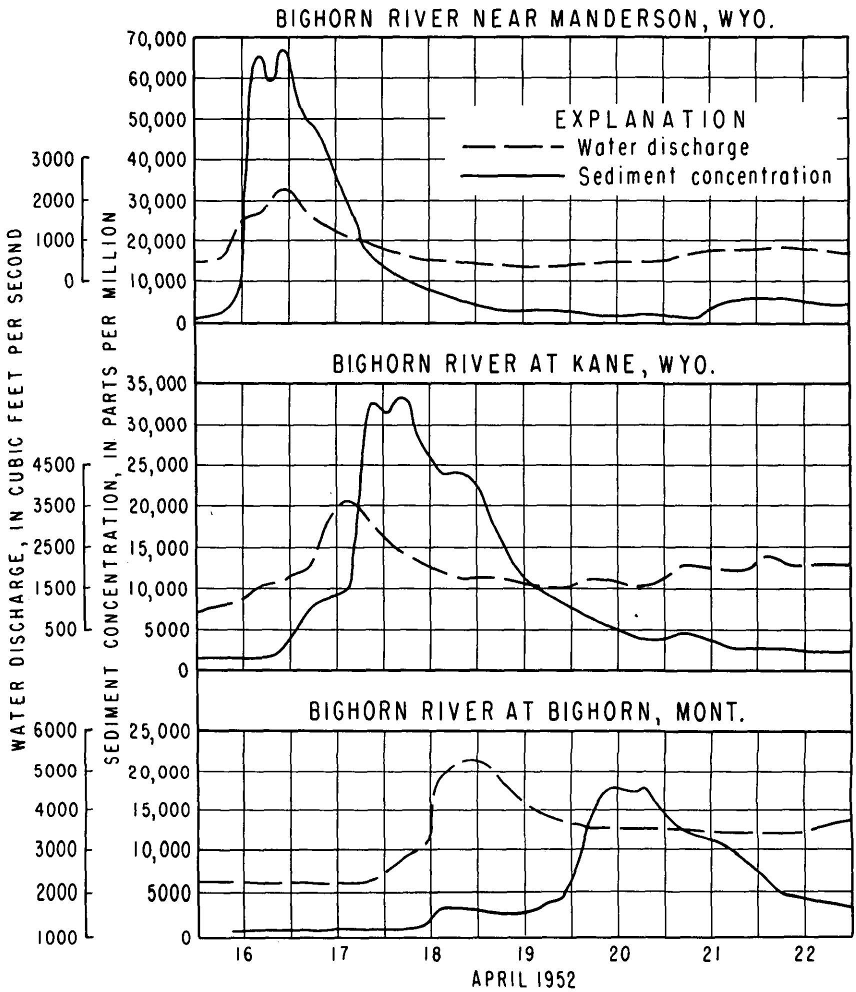

FIGURE 11.—Discharge and sediment concentration of Bighorn River, April 16–22, 1952 (Heidel 1956).

Manderson, Wyo., during the rise in April 1952 was mostly water that had been in the channel before the storm and contained little fine sediment. The increase in sediment discharge at Bighorn at about noon on April 18 resulted from erosion of the channel bed and banks. The sediment discharged after about noon on April 19 at Bighorn was mostly fine sediment that was eroded far upstream during the storm runoff on April 16 and that caused peaks of sediment concentration near Manderson late on April 16 and at Kane during the night of April 17–18. The difference in source of sediments at the time of the peak flow, as compared with that at the time of the peak sediment concentration for certain storms on a stream like the Bighorn River, might be significant in studies of waste dispersal.

DEPOSITION

Sediment is deposited from streamflow in accordance with the same physical laws that are involved in sediment transportation, but sediment deposition can, perhaps, be better explained through a generalized statement of the movement of sediment particles within a flow and their settling from a flow.

Sediment particles, being more dense than water, tend to settle with respect to the water at rates that depend on the differences in density between the particles and the water, on the viscosity of the water, and on the size, shape, and flocculation of the particles. Water temperature and the presence of interfering sediment particles influence the effective viscosity of the fluid-sediment mixture and hence the fall velocities of the particles. Also, any flocculation of the fine sediment changes the fall velocities of the affected particles. The amount of flocculation depends on the size, shape, and composition of the particles and on the turbulence and chemical composition of the water. In still water and at concentrations low enough to prevent significant interference among the particles, each particle settles at its own characteristic fall velocity in a particular fluid of given density and viscosity.

In turbulent flow, water moves up and down and laterally with respect to its general direction of flow. If a particular volume of water in turbulent flow could be identified, the components of that volume of water at the end of an elapsed time would be scattered in the flow. On the average, at the end of the elapsed time the components would be downstream a distance determined by the time and the velocity of flow; they would be at a lower average elevation that could be computed from the distance of downstream travel and the slope of the stream.

Similarly, sediment particles in a particular volume of flow at a particular time will be dispersed from that volume by turbulent flow.

After a period of time, some particles will be higher in the flow and some will be lower than the position of the original volume of water. On the average, the sediment particles will be lower in the flow than the components of the water from the same original volume by distances that depend on the fall velocities of the particles. If none of the particles reach the streambed and secondary effects are disregarded, the difference between the average elevation of the sediment particles of a particular fall velocity and the average elevation of the water originally in the same volume with these particles equals the product of the elapsed time and the fall velocity. For example, after an elapsed time of 100 seconds, sand grains of about 0.5 mm might be on the average 20 feet below the water with which they started, whereas silt grains of 0.02 mm might be on the average only 0.1 foot lower than the water with which they started. Even fine clay particles settle, but very slowly, with respect to the surrounding water. Thus, the individual sediment particles on the average are continually settling toward the streambed, even while a fraction of them are moving upward in the flow. Eventually, all the particles in the turbulent flow would settle out if no additional particles entered the flow and if those particles that settled to the streambed remained there.

When water flows over unconsolidated sediment at high enough velocities, some sediment particles are removed from the bed. Of those that are lifted or started into motion, some fall back to the bed but some are carried upward. Even though the number that move upward is only a small fraction of the total number that are shifted at the bed, the ones that do escape upward are added to the particles in suspension. If during a particular time the quantity of these particles that escape upward from the bed into suspension is less than the quantity that settles from suspension to the bed, net deposition occurs. Although no net deposition occurs, individual particles are continually being interchanged between the bed and suspension in the fluid. Because of this continual interchange, a slight decrease in transporting ability of the flow immediately shifts the balance between particles arriving at the bed and those leaving the bed and may quickly cause net deposition.

The balance between sediment particles leaving the bed and those arriving at the bed depends on the local ability of the flow to transport sediment rather than on the average parameters of flow at a cross section. This fact is shown by scour around the upstream corners of an obstruction and by fill downstream from an obstruction such as a boulder on the streambed. Hence, deposition of sediment may be caused by either a general or a very local reduction in transporting ability. Either the local or the general deposition may, for example, trap radioactive sediments.

Sediment deposits can be classified, with some overlapping, into those associated with lakes or reservoirs, tidal areas, stream channels, or flood plains.

LAKE AND RESERVOIR DEPOSITS

Most natural lakes collect some fine sediment from inflowing water or from aeolian sediments that reach the lake surface. The fine sediment settles to the lakebed, where it usually accumulates below the effect of wave action or in places protected from wave action. Except for some of the sediment deposited in deltas at the mouths of streams that enter lakes, most lake sediments are in relatively permanent deposits that are unlikely to be appreciably dispersed by later movement.

When a stream that carries sediment nears a lake or reservoir, it begins to drop its coarse sediment as soon as the velocity slows down even slightly unless the stream was not transporting its capacity load of coarse sediment. The area where the stream begins to lose its transporting ability can be approximated from a computation to determine the upstream end of the backwater curve. Further slowing of the flow as it approaches the reservoir causes additional deposition. Not all the coarse sediment is deposited first, but a relatively much larger percentage of the coarse sediment than of the fine sediment is deposited where the flow begins to slow down. As the velocity decreases more and more, even the silt and clay deposit in appreciable quantity.

Some fine sediment may pass through the delta area at the head of a reservoir without being deposited. If the inflowing water-sediment mixture is less dense than the water of the reservoir because of chemical composition or temperature, the inflowing mixture may spread out on the surface of the reservoir water and maintain a separate identity for a considerable distance. Occasionally, the incoming water-sediment mixture may move into the reservoir as a distinct stream that is somewhere between the surface and the bed of the reservoir. If the inflow is more dense than the reservoir water, the incoming stream may flow by the action of gravity along the downstream face of the delta and toward the lowest place in the reservoir, which is usually the upstream face of the dam. However, the dense incoming stream may lose its identity before reaching the lowest point in a large reservoir. According to Gould (1951, p. 48), nearly half the tonnage of sediment inflow to Lake Mead from 1935 to 1948, inclusive, was deposited in the deep parts of the reservoir beyond the delta areas.

The shape of the delta that is formed by the deposited sediments varies with channel shape, fall velocities of the particles, and fluctuations of the lake or reservoir surface. Generally, the depth of a delta

deposit in a stream channel increases slowly downstream to a maximum and then decreases rather abruptly on the downstream face of the delta. For a simple condition of constant surface elevation of a reservoir and well-defined entrance channel, the location and, to some degree, the form of the reservoir deposit can be approximated by methods described by Fowler (1957).

A major change in elevation of a lake or reservoir may shift the location of the delta of an inflowing stream. A delta formed at one elevation of the downstream water surface may be reworked at a lower elevation of the water surface, or it may be partly covered by a delta that is deposited farther upstream at a higher elevation of the reservoir surface. Sediment exposed during low stages of a reservoir usually compacts faster and becomes more resistant to later movement than sediment that is always submerged.

The effect of delta deposits can extend far up some streams that have flat slopes. After the original stream channel has been filled at the approach to the reservoir or lake, the delta gradually widens to cover the valley floor. Vegetation may grow on parts of the delta and retard the flow. Sometimes a wide valley floor, such as the floor of the Rio Grande valley above Elephant Butte Reservoir in New Mexico, may become so choked with vegetation growing in a swampy area that all sediment during low flows and all except a part of the very fine sediment during high flows may be deposited on the valley floor. The large sediment deposits reduce the slopes of the stream channel and the valley floor and may also raise the water table and increase growth of vegetation. The decrease of slope may work progressively upstream. Permanent sediment deposits may be formed both in the channel and on the valley floor wherever the slope is flattened. (Of course, a delta that starts the choking of a channel may be formed where a major tributary that transports large quantities of sediment enters a broad valley of a larger stream as well as at the entrance to a reservoir.)

Many small reservoirs retard flow sufficiently to trap all incoming coarse sediment and part of the fine sediment. The coarse sediment is deposited as a delta that usually extends higher than the spillway elevation. Delta formations generally, however, extend only short distances upstream from small reservoirs because of the steepness of the slopes of the incoming channels. Fine sediment temporarily deposited in a small reservoir may later be flushed from it by succeeding flows. The trap efficiency of a reservoir (Brune, 1953) is a function mainly of the fall velocities of the sediment particles and of the degree to which the water is slowed as it passes through the reservoir. The amount of flocculation of the fine particles frequently has a major effect on trap efficiency because it affects the fall velocities of the silt

and clay, which are the sediments that otherwise might pass through the reservoir. Many small reservoirs fill rapidly with sediment, so that their trap efficiencies usually decrease with time.

Trap efficiencies may be high for reservoirs on small drainage areas. H. P. Guy and others (written communication, 1958) reported trap efficiencies of 50 percent for one reservoir and 80 to 98 percent for eight others than were included in a cooperative study by the U.S. Soil Conservation Service and the Geological Survey. One of the nine reservoirs had a drainage area of 47.3 square miles and the other eight, including the one with the low trap efficiency, had drainage areas of less than 5 square miles.

TIDEWATER DEPOSITS

The factors controlling sediment deposition in channels whose flow is affected by tides differ from the factors controlling deposition at the entrance to lakes and reservoirs in two significant ways. One way is that the surface level of the ocean varies with the tides; the other is that the chemical composition of the ocean water is much different from that of the inflowing fresh water.

The changes in tide affect the place of deposition of the stream sediments. A stream transporting some sediment sizes at its full ability will begin to deposit some particles of these sizes where the stream is first slowed by the effect of the ocean level. The place of this first deposition may vary several miles and depends on whether the flow is affected by high or low tide. Thus along an appreciable reach of tidal stream, sediment deposition may be intermittent. Also, some sediment deposited at high tides may be eroded from the streambed at low tides. Farther downstream, sediment may deposit slowly at low tide and much faster at high tide. Of course, the amount and place of deposition of sediment also vary with the discharge of the stream. If the streamflow is low, some fine sediment may even be carried back upstream while the tide is rising and be deposited before downstream flow begins again. Especially during floods, some fine sediments may be carried far out into an ocean or bay by the stream current.

The fine sediments usually flocculate readily when they meet the saline water near the mouth of a stream that enters an ocean or a salt-water bay. The flocculated particles then settle faster with respect to the flow than less flocculated particles of the same discrete sizes. The upstream extent of the salt-water intrusion in a river or estuary varies within each tidal cycle and with the flow of the stream and the shape of the channel (Keighton, 1954).

The water-sediment mixture that enters salt water from a river is practically always less dense than the salt water, and hence the

mixture may spread out over the salt water but probably seldom moves under it in the form of a density current that might carry large amounts of fine sediment far from the river mouth. Also, the effects of the tide may help to mix saline water with the river water and increase the amount of flocculation of the fine sediment. These facts do not mean that vast tonnages of fine sediment are not carried beyond the mouths of tidal rivers, but they do mean that the proportion thus transported would probably be far greater if the stream entered a fresh-water reservoir.

The deltas formed at the mouths of large rivers that enter large saline bodies of water—such as the delta at the mouth of the Mississippi River—may be very extensive. Sediment coming down the Mississippi River may be deposited in any one of several channels, lakes, or marshy areas. Deltas of streams smaller than the Mississippi River may also be complex in form and have a wide areal range over which sediment may be deposited temporarily or permanently. To a considerable extent, each tidal delta is an individual problem in sediment deposition.

The effects of tide and saline water on sediment deposition near the mouths of tidal streams are so complex and variable that they can seldom be evaluated satisfactorily without extensive investigation of each stream. The probability of temporary, semipermanent, or permanent deposition of much of even the finest sediments—which are the most significant ones for some waste-disposal problems—should be considered carefully for the whole tidal reach of each stream.

CHANNEL DEPOSITS

Individual particles of sediment may be deposited within a stream channel because of a local or a general reduction of transporting ability of the stream or because suspended-sediment particles interchange with particles on the bed even though no net deposition occurs.

Sediment is deposited where velocity and turbulence are locally reduced downstream from rocks or stalled driftwood or at pockets along the streambank. Many such deposits may be so small or temporary or both that they would not be significant for many fluvial-sediment problems. Local deposition also often occurs during decreasing flow at constricted stream sections such as those at many bridges. If the bed of the stream scours during high flow through the constriction, the scour hole generally refills partly or completely when the flow decreases. Some sediment in the fill may come from the streambed upstream, and some from recent sheet erosion far upstream. Usually, local fills at channel constrictions are partly or entirely removed by the next flood.

Many streams have alternate pools and riffles and, in general, an inadequate supply of available sands at least during high flows. Hence, the pools generally scour during sustained high flows and fill during sustained low flows. A sediment deposit several feet deep may be removed from a pool or accumulated in a pool over a period of time (Leopold and Maddock, 1953, p. 32) even though the elevation of the riffle downstream from the pool did not change. Thus large local deposits, often temporary, may be formed in pools when the stream is neither generally aggrading nor degrading.

General deposits of sediment in a channel occur when more sediment is brought to a reach of stream than can be transported through the reach by within-bank flows. A reach of stream that has been in approximate equilibrium with its sediment load may have its transporting ability decreased by weed, brush, or tree growth on the banks or within the channel, so that the channel gradually aggrades along the reach. Accelerated erosion of coarse sediment upstream may increase the discharge of bed material into the upper end of a reach to a rate appreciably greater than the ability of the stream to transport bed material and thus cause general deposition of sediment. As a channel fills, its transporting ability usually decreases further and thereby increases the rate of deposition.

Individual sediment particles deposit in the bed of a uniform channel while the average bed elevation is stable and no net deposition is occurring. As was mentioned previously, even the smallest sediment particles settle to the streambed occasionally, and the large particles reach the bed relatively more often. According to D. B. Simons (oral communication, 1960), the concentration of clay particles in the water in the pore spaces of a sand bed is about the same as their concentration in the flow above the bed. While on the bed, either fine or coarse particles recently in suspension may become buried, especially if they happened to land in a depression between sand dunes. Considerable shifting and mixing of particles on the bed also occur for other types of bed configuration than dunes. Simons and Richardson (written communication, 1960) reported that fine particles introduced in suspension in a flume were sometimes later found in noticeable quantities well below the surface of the sand bed in the flume.

DEPOSITS ON FLOOD PLAINS

Flood plains are built from fluvial deposits that accumulate over long periods of time. The sediment may be deposited either within a channel as it shifts laterally or on the surface of a flood plain. Because channel deposits have been mentioned previously, only the deposits on the surface of flood plains are discussed here.

Flood plains generally are submerged only at times when the flow and sediment concentration are high. On the other hand, the water that leaves a channel to spread out over a flood plain usually comes from the upper layers of flow of the main stream and hence has a lower concentration, particularly of relatively coarse sediment, than the average for all layers of flow. However, at cutoffs and levee breaks, coarse sediments in large quantity may be transported from the channel onto the flood plain.

The relatively low velocities of flow over flood plains indicate low transporting ability of the flood-plain flows as compared with that of the main-channel flows. Hence, some or most sediment in water that flows from the main channel onto the flood plain may be deposited. The coarsest part of the load may be dropped close to the edge of the main channel in the form of natural levees. Much of the remaining sediment may deposit on the flood plain, sometimes, in a relatively uniform layer and sometimes mainly in those areas where vegetation or other obstructions are especially effective in retarding the flow.

Of course, flood-plain deposits may in part be picked up later by meandering of the main channel during flood lows or by erosion on the surface of the flood plain; the erosion may be by tributary inflow or by overbank flow on the flood plain. However, in general, most flood-plain deposits are relatively permanent. Any recently deposited materials on a flood plain are likely to be near the surface and generally widely dispersed in thin layers over the flood plain.

Many flood plains contain swampy areas or oxbow lakes that may trap most of the incoming sediment. Sediment deposits in marshes are usually permanent, but sediment deposits in flood-plain lakes may be at least partly removed by later high flows.

METHODS OF MEASURING SEDIMENT DISCHARGE

Methods for computing the discharge of sediment; especially the fine sediment, generally are laborious and rather inexact. At the present time and probably for many years to come, the discharge of fine sediments, and perhaps of most coarse sediments as well, can be determined better by sediment sampling than by theoretical computations. Hence, some methods commonly used for measuring discharge of total sediment, suspended sediment, and bedload are explained here briefly.

TOTAL SEDIMENT

One method of measuring the total sediment discharge of a stream is to collect all the sediment, and sometimes all the water, for a known time and then determine the volume and the specific weight of sediment that has accumulated. Sediment accumulations in reservoirs are frequently used to indicate average sediment discharges from the

tributary areas, often over long periods of time. Corrections are applied, if necessary, for sediment that passes through the reservoir (Brune, 1953).

The total sediment load of a few small streams in the laboratory or field can be caught in tanks during individual storms or periods of time. If the sediment accumulations in the tanks are carefully measured, this method is probably the most accurate one for determining total sediment discharges.

Some streams and flumes have free fall from the end of a conduit or flume, and slot or nozzle samplers may sometimes be moved manually through the nape of the discharge to obtain samples from which the total sediment discharge at the sampling times can be computed. A few samplers such as the Coshocton sampler (Parsons, 1955) collect a known fraction of the flow and sediment discharge continually during a period of runoff. These samplers are generally limited at the present time to the sampling of small flows.

In natural streams of reasonably large size, a natural constriction (Colby and Hembree, 1955) or artificially induced turbulence (Benedict, and others, 1955) may sometimes suspend practically all the sediment discharge, so that it can be measured as suspended-sediment discharge.

SUSPENDED SEDIMENT

In the United States, most suspended-sediment samples are now collected with the U.S. DH-48, U.S. D-43, or U.S. D-49 sediment samplers, which were developed and described by the Federal Inter-Agency River Basin Committee (1952a). The first of these samplers is for suspension on a rod and is usually used when a stream is wadable. The other two are for suspension on cables that are wound on reels; they are usually used from bridges or cableways. Each of the three is a nozzle-type sampler that is designed to collect water-sediment mixture whenever the nozzle is submerged and at an intake velocity that approximates the horizontal velocity of the stream at the nozzle intake. The flow is neither appreciably retarded nor accelerated at the intake, and the sediment particles do not migrate unnaturally with respect to the surrounding water while the sample enters the intake.

These samplers generally are lowered and raised at a constant rate while they collect a depth-integrated sample of water-sediment mixture from the stream surface to within a few tenths of a foot of the streambed. The collected samples, therefore, have volumes that are theoretically proportional to the water discharge per unit width at the traversed vertical.

Samples are usually collected at each of several verticals across the stream. If the verticals are uniformly spaced and the sampler is

moved at a constant vertical rate during sampling at all verticals, a velocity-weighted concentration for the whole sampled cross section can be computed from the ratio by weight of the total sediment collected to the total water-sediment mixture collected. If the verticals are laterally spaced to represent equal volumes of water discharge, the concentration is separately determined for the samples at each vertical, and the concentrations for all verticals are averaged to obtain the velocity-weighted concentration for the entire cross section. The average velocity-weighted concentration for a cross section can be multiplied by the water discharge and by a conversion constant to compute the measured discharge of suspended sediment. A velocity-weighted concentration for a vertical or for an entire cross section can be analyzed to determine the relative concentrations of the different particle sizes.

The U.S. P–46 sampler also was developed by the Federal Inter-Agency River Basin Committee (1952a). It has a valve that can be opened and closed by remote control from a bridge or cableway so that water-sediment mixture enters the sampler only during controllable periods. It is used to obtain sediment samples at a point in a vertical. It is also used for collecting depth-integrated samples during sampler movement in only one direction at a vertical, rather than both up and down, and for sampling only part of the vertical depth of deep and swift flows.

Individual samples of coarse sediment in suspension are generally much less dependable measures of average concentrations in a stream than individual samples of fine sediment. The concentrations of fine sediment usually are about the same throughout a cross section at a particular time, but the coarse sediment is less uniformly distributed both laterally and vertically. Also, measured concentrations of coarse sediment show short-time fluctuations because of pulsations in velocity, "boils," and possible pickup of sediment from the streambed.

BEDLOAD

Many devices have been used or have been suggested for measuring the bedload discharge of streams. The usual type of bedload sampler is attached to a cable and then lowered to the streambed. There, it either screens the coarse sediment from the water-sediment mixture near the bed or else traps the coarse sediment near the bed between baffles or in pockets, where the velocity is reduced within the sampler. Most bedload samplers have been described by D. W. Hubbell (1963–64). Sampling with these devices has shown that the bedload discharge is an appreciable amount in many streams and that it is highly variable with time and location on the streambed (Einstein, 1937). Studies in field streams have helped support laboratory

findings to some extent, but none of the bedload samplers has yet been used widely in the United States.

SEDIMENT-DISCHARGE RECORDS

The general methods applied in computing sediment discharges depend on the kind, accuracy, and frequency of the available samples and on the expected use of the computed discharges. Frequently, a combination of different methods is used. The general procedure differs enough to be discussed separately for computations that are based on frequent sediment-discharge measurements and for those based on occasional sediment-discharge measurements.

For this discussion, the available sediment samples are assumed to be representative of the velocity-weighted concentration at the time of sampling or to have been adjusted so that they are representative.

Some idea of sampling coverage, sources of available sediment records, and the variability of concentration of fluvial sediments in the United States can be given briefly; however, studies for special problems such as the disposal of radioactive wastes in a stream will usually require extensive local investigation of the fluvial sediments

FROM FREQUENT SEDIMENT-DISCHARGE MEASUREMENTS

Sediment samples are generally obtained infrequently unless a reasonably accurate time distribution of the sediment concentration and discharge is desired in addition to the average sediment discharges for a runoff event, for months, or for longer periods of time. The time distribution can be indicated by plotting the concentrations of the sediment samples on a hydrograph of gage height or water discharge and drawing a continuous curve through the concentrations. Between the times of sampling, the continuous curve is based on all pertinent information and on the judgment of the computer. The information usually includes a graph of average sediment discharge or concentration plotted against gage height or streamflow, or else some typical sediment-concentration graphs plotted against time during periods of storm runoff. It may include data on storm distribution and intensity, stream velocity, soil condition, or vegetal cover.

If the flow and sediment concentration are reasonably constant during a day, the daily average concentration, the daily average flow, and a conversion constant are multiplied together to obtain the sediment discharge in weight per unit of time. If the flow and concentration change widely or if sediment discharge is needed for parts of a day, the computation of sediment discharge is based on average concentration from the continuous curve and on average

flow for a few minutes or hours during which neither flow nor concentration varied widely.

FROM OCCASIONAL SEDIMENT-DISCHARGE MEASUREMENTS

When only occasional samples are available, the drawing of a contInous curve of sediment concentration may be impracticable. However, if a few dozen occasional measurements of sediment discharge have been made over the usual range of flow of a stream, the sediment discharges or concentrations can be plotted against the streamflows at the sampling times to define an average sediment discharge or sediment concentration curve for a cross section. Usually, individual points scatter widely (fig. 6) from such a curve. Separate curves may be prepared for fine-material load and for bed-material load if separate discharges of the two are required or if different adjustments are to be applied to each average curve. The adjustments from the average curves might be for season of the year or for source of the sediment in the drainage basin if the sediment is mainly fine; they might be for water temperature or variations in mean stream velocity if the sediment is mostly coarse.

For a particular sediment station, one or more adjusted or unadjusted curves based on sediment-discharge measurements and concurrent streamflows can be used to compute roughly the sediment discharge at any time when the streamflow is known. Although sediment discharges computed from such sediment-discharge curves may be reasonably correct on the average, at individual times they will be greatly in error. Usually, these curves should not be used to compute daily sediment discharges directly from daily average rates of flow.

If the concentrations of individual samples or sets of samples collected at one time correctly represent the sediment concentrations at the times of sampling, more accurate sediment discharges generally can be obtained if the sediment-discharge curve is adjusted on the basis of each sediment-discharge measurement that is computed from such samples. This adjustment may be made by shifting the curve in a manner comparable with shifting a curve of stage-discharge relation to individual streamflow measurements. The adjustment can probably be made more simply by drawing a hydrograph of the sediment discharges computed from the average sediment-discharge curve and other average relations and then shifting this hydrograph by eye to make it pass through the instantaneous sediment discharge at the time of each sampling.

Average sediment discharges for a day, storm period, month, year, or longer period can be computed directly from a flow-duration curve for the period and a sediment-discharge curve that is assumed to be

representative for the period. The range of flow is divided into several subranges, and the fraction of the total time that the flow was within each subrange is determined from the flow-duration curve. The sediment discharge for the midpoint or other representative flow for each subrange is determined from the sediment-discharge curve. The average sediment discharge during the period equals the sum of the products of sediment discharge and fraction of time during which each sediment discharge applied.

Theoretically, the duration curve and the sediment-discharge curve should each be based on flows for the same length of time. That is, if the sediment-discharge curve is based on a relation between sediment discharge and water discharge for a short period of time while the samples were being collected, the duration curve should also be based on average flows for a comparably short period of time. If the sediment-discharge curve is based on the relation between average sediment discharge and average water discharge for a day, the duration curve should be based on average daily flows. The practical importance of this theoretical restriction depends on the flashiness of flow and concentration, particularly at the higher rates of flow for the stream.

RECORDS IN THE UNITED STATES

Records of sediment discharge in streams in the United States are inadequate for many studies. According to Langbein and Hoyt (1959, p. 31), daily records of suspended-sediment discharge were obtained during 1956 at only about 220 gaging stations, of which about 150 were operated by the U.S. Geological Survey and about 50 by the U.S. Army Corps of Engineers. Other Federal agencies that obtain sediment records at stream stations or determine sediment accumulations in reservoirs include the Bureau of Reclamation, Soil Conservation Service, Agricultural Research Service, Forest Service, International Boundary Commission, and Tennessee Valley Authority. Sediment losses from small areas are obtained by some of the above agencies and by many agricultural experiment stations of colleges and universities. Most sediment records of the U.S. Geological Survey are published annually in water-supply papers under the general title "Quality of Surface Waters of the United States."

Inventories of published and unpublished records of sediment discharge in the United States before October 1, 1950, have been prepared in two bulletins for limited distribution and administrative use by the Subcommittee on Sedimentation of the Federal Inter-Agency River Basin Committee (1949; 1952b). The bulletins contain concise lists of station locations, drainage areas, periods of record, some information on types of samplers and sampling methods, and references or sources of the tabulated data.

Some size analyses of sediment samples are usually available if sediment discharges have been computed, but these size analyses or part of them may not have been released with the records of sediment discharge. In general, more detailed sediment information is contained in an agency's files than is published. The unpublished information for a river basin or for a State may include a listing of sediment-discharge records that are available in the files of other agencies.

The accuracy of available sediment records is variable both with time and from station to station. Often, some idea of the general accuracy of records at a sediment station can be based on the flashiness of the stream, frequency of sampling, and kinds of equipment that were used, but the experience and ability of the computers of the records can seldom be evaluated satisfactorily.

Some inaccuracies in sediment records are compensating over periods of time, but a noncompensating error in most sediment records results from the fact that samples usually represent velocity-weighted sediment concentrations of the flow exclusive of the bottom 0.3 or 0.4 foot of depth at each sampling vertical. The difference between measured or computed sediment discharge and total sediment discharge of a stream consists of bedload discharge and part of the suspended-sediment discharge near the streambed where concentration of the suspended sediment, especially the coarse sediment, is highest. As the difference is largely coarse sediment, it is significant in most studies of channel behavior but is of comparatively lesser importance in studies involving fine material.

The amount of sediment discharge that is usually not measured in sand-bed streams is a function mainly of mean velocity. A rough idea of the amount for sand-bed streams is given by studies for Niobrara River near Cody, Nebr., which showed about 4.5 and 21 tons per day per foot of stream width for mean velocities of 2.5 and 4.0 feet per second, respectively. These quantities are roughly half of the total discharge of sands of Niobrara River near Cody for the same mean velocities.

Average sediment concentrations differ widely from place to place within the United States. The annual average concentrations computed from the ratio of annual sediment discharge to annual water discharge for the water year ending September 30, 1955, were about 42, 75, 7,200, and 140,000 ppm, for the Mohawk River at Cohoes, N.Y., the Scantic River at Broad Brook, Conn., the Rio Chama near Abiquiu, N. Mex., and the Rio Puerco near Bernardo, N. Mex., respectively. A generalized map for streams in the United States east

of the Mississippi River (Rainwater, 1962) shows approximate average concentrations of sediment weighted with water discharge. (See fig. 12.) A comparable map for streams west of the Mississippi River would show much wider differences in average concentrations within short distances. In general, the sediment concentrations in mountain streams are very low. The concentrations in streams draining the plateau and plains regions in the western United States tend to be high, sometimes very high.

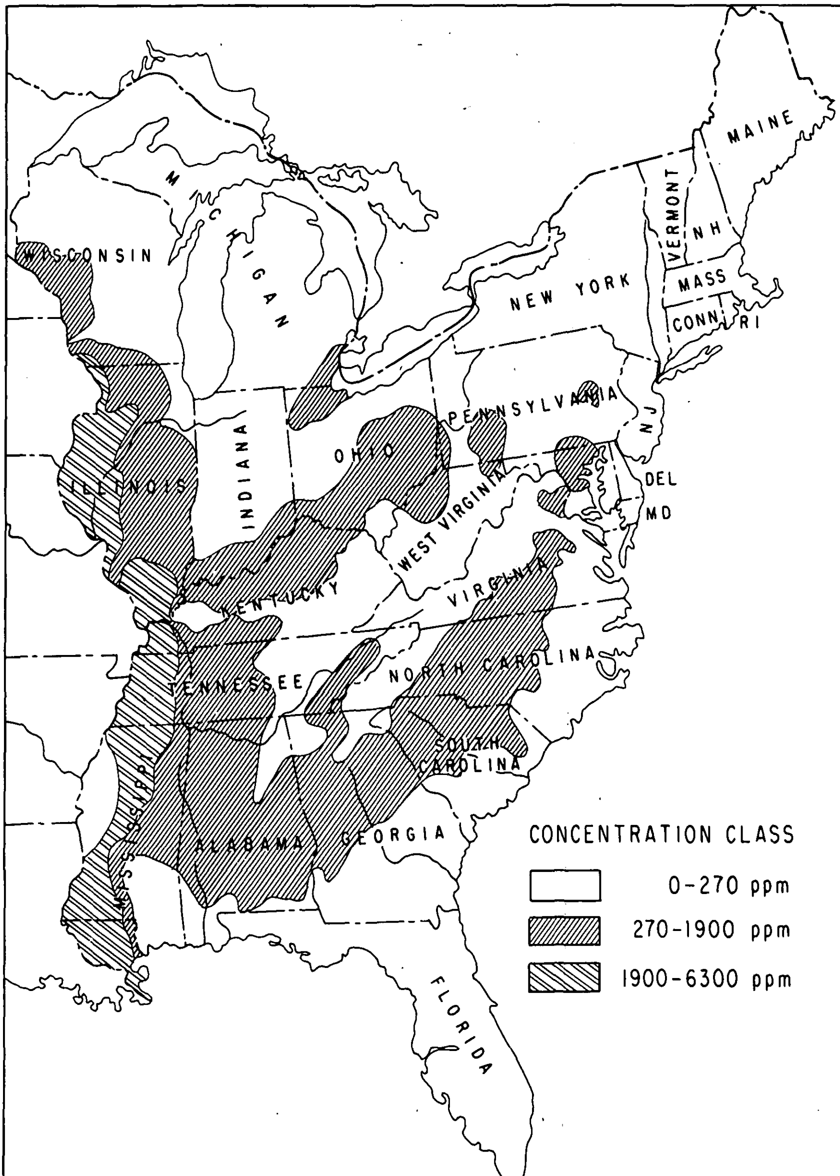

FIGURE 12.—Map of eastern United States showing concentration of suspended sediment in streams (Rainwater, 1962).

REFERENCES

Bagnold, R. A., 1956, The flow of cohesionless grains in fluids: Royal Soc. [London] Philos. Trans., v. 249, p. 235–297.

Benedict, P. C., Albertson, M. L., and Matejka, D. Q., 1955, Total sediment load measured in turbulence flume: Am. Soc. Civil Engineers Trans., v. 120, p. 457–484.

Brooks, N. H., 1958, Mechanics of streams with movable beds of fine sand: Am. Soc. Civil Engineers Trans., v. 123, p. 526–594.

Brune, G. M., 1953, Trap efficiency of reservoirs: Am. Geophys. Union Trans., v. 34, p. 407–418.

Chepil, W. S., and Woodruff, N. P., 1957, Sedimentary characteristics of dust storms; II, Visibility and dust concentration: Am. Jour. Sci., v. 255, p. 104–114.

Colby, B. R., 1956, Relationship of sediment discharge to streamflow: U.S. Geol. Survey open-file report.

Colby, B. R., and Hembree, C. H., 1955, Computations of total sediment discharge Niobrara River near Cody, Nebraska: U.S. Geol. Survey Water-Supply Paper 1357, 187 p.

Einstein, H. A., 1937, Die Eichung des im Rhein verwendeten Geschiebefängers [Calibrating the bed-load trap as used in the Rhein]: Schweizer. Bauzeitung, v. 110, no. 14 [in German]. Translations on file at U.S. Soil Conserv. Service, Washington, D.C.; at California Inst. Technology, Pasadena (translation 39–7); and at U.S. Waterways Expt. Sta., Vicksburg, Miss. (translation by Ott, W. P.).

——— 1950, The bed-load function for sediment transportation in open channel flows: U.S. Dept. Agriculture Tech. Bull. 1026, 70 p.

Einstein, H. A., Anderson, A. G., and Johnson, J. W., 1940, A distinction between bed-load and suspended load in natural streams: Am. Geophys. Union Trans., pt. 2, p. 628–633.

Federal Inter-Agency River Basin Committee, 1949, Inventory of published and unpublished sediment-load data in the United States: Subcomm. on Sedimentation Bull. 1, 82 p.

——— 1952a, The design of improved types of suspended sediment samplers, *in* A study of methods used in measurement and analysis of sediment loads in streams: Rept. 6, 103 p.

——— 1952b, Inventory of published and unpublished sediment-load data in the United States: Subcomm. on Sedimentation, Supp.—1946 to 1950, Bull. 4, p. 49.

Fowler, L. C., 1957, Determination of location and rate of growth of delta formations: U.S. Army Engineer Div., Missouri River, Sediment Memo. 6, 13 p.

Gould, H. R., 1951, Some quantitative aspects of Lake Mead turbidity currents, *in* Soc. Econ. Paleontologists and Mineralogists, Turbidity currents and the transportation of coarse sediments to deep water—a symposium: Soc. Econ. Paleontologists and Mineralogists Spec. Pub. 2, p. 34–52.

Heidel, S. G., 1956, The progressive lag of sediment concentration with flood waves: Am. Geophys. Union Trans., v. 37, pt. 1, p. 56–66.

Hubbell, D. W., 1963, Apparatus and techniques for measuring bed-load: U.S. Geol. Survey Water-Supply Paper 1748 (in press).

Hubbell, D. W., and Matejka, D. Q., 1959, Investigations of sediment transportation, Middle Loup River at Dunning, Nebraska, with application of data from turbulence flume: U.S. Geol. Survey Water-Supply Paper 1476, 123 p.

Keighton, W. B., 1954, The investigation of chemical quality of water in tidal rivers: U.S. Geol. Survey open-file report.

Keulegan, G. H., 1938, Laws of turbulent flow in open channels: U.S. Natl. Bur. Standards, Jour. Research, v. 21, no. 6, p. 707–741.

Lane, E. W., Chairman, and others, 1947, Report of the Subcommittee on Sediment Terminology: Am. Geophys. Union Trans., v. 28, no. 6, p. 936–938.

Langbein, W. B., and Hoyt, W. G., 1959, Water facts: New York, Ronald Press Co., 288 p.

Leopold, L. B., and Maddock, Thomas, Jr., 1953, The hydraulic geometry of stream channels and some physiographic implications: U.S. Geol. Survey Prof. Paper 252, 56 p.

Meyer-Peter, E., and Müller, R., 1948, Formulas for bed-load transport: Internat. Assoc. Hydraulic Structures Research, 2d Mtg., Stockholm, Sweden, p. 39–64.

Musgrave, G. W., 1947, The quantitative evaluation of factors in water erosion, a first approximation: Jour. Soil and Water Conserv., v. 2, p. 133–138.

Parsons, D. A., 1955, Coshocton-type runoff samplers: U.S. Dept. Agriculture, Agr. Research Service, ARS–41–2, 9 p.

Rainwater, F. H., 1962, Stream composition of the conterminous United States: U.S. Geol. Survey Hydrol. Atlas 61.

Rouse, Hunter, 1937, Modern conceptions of the mechanics of fluid turbulence: Am. Soc. Civil Engineers Trans., v. 102, p. 463–543.

Sayre, W. W., Guy, H. P., and Chamberlain, A. R., 1963, Uptake and transport of radionuclides by stream sediments: U.S. Geol. Survey Prof. Paper 433–A, 35 p.

Straub, L. G., 1954, Terminal report on transportation characteristics of Missouri River sediment: Minnesota Univ. Hydraulic Lab., 24 p.

Twenhofel, W. H., 1939, Principles of sedimentation: New York and London, McGraw-Hill Book Co., 610 p.

Vanoni, V. A., and Brooks, N. H., 1957, Laboratory studies of the roughness and suspended load of alluvial streams: California Inst. Technology Sediment Lab. Rept. E–68, 121 p.

Woodburn, Russell, 1949, Science studies a gully: U.S. Dept. Agriculture Soil Conserv., v. 11, no. 1, p. 11–13, 22.

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