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Techniques of Water-Resources Investigations
of the United States Geological Survey

Chapter C1

FLUVIAL SEDIMENT CONCEPTS

By Harold P. Guy

Book 3
APPLICATIONS OF HYDRAULICS

fore row crops are planted and left on the surface when small grain is seeded.

The values of the variables of the equation are obtianed as follows: the rainfall factor, R, for west central Indiana [fig. 5] is 185. The factor K is a measure of the erodibility of a given soil and is evaluated independently of the effects of topography LS, cover and management C, and supplementary practices P. When those conditions of independence are met and LSCP=1, K equals E/R or 0.38 ton per unit of erosion index for Russell silt loam. For an 8-percent 200-foot slope, the topographic factor, LS, is found to be 1.41 [fig. 6].

The cropping factor, C, is computed by crop stages for the entire 4-year period. The input for calculation of C includes average planting and harvesting dates, productivity, disposition of crop residues, tillage, and distribution curves of the erosion index throughout the year. The ratio of soil loss from cropland corresponding loss from continous fallow, by each crop stage, is found in voluminous tables in Agricultural Handbook 282 [Wischmeier and Smith, 1965]. The value of C for central Indiana is computed to be 0.119. The practice factor, P=0.6, is based on the decision to contour and depends upon land slope and slope length according to criteria given in Handbook 282. The average annual soil-loss rate for this Indiana field would be expected to be E=(185)(0.38)(1.41)(0.119)(0.6)=7.1 tons/acre.

In the above example, if the conservation practice of stripcropping with alternate meadows were used, P would be 0.3 and E would then be 3.5 instead of 7.1 tons per acre. Also, if minimum tillage of corn were combined with contour planting, the cropping factor, C, would be 0.075 instead of 0.119, and with the use of alternate meadows (P=0.3), E would be 2.2 tons per acre. It is, therefore, most evident that land use is a very significant element in the amount of sediment eroded from a given environmental complex.

Vice, Guy, and Ferguson (1969) estimated the gross erosion in a basin undergoing extensive highway construction through consideration of the amount and size of material transported by the stream from the basin and the size of the residual and eroding sediments in the basin. The assumption was made that all the eroded clay found its way through the channel system and hence was measured as basin output. The amount of eroded sand- and silt-sized materials could then be determined by direct proportions from the percentages of clay, silt, and sand in both the soils and sediment transported from the basin.

Predicting gully erosion

Gullies, or deep and steep-walled upland channels, are commonly associated with a concentration of flow over areas of deep friable subsoils where valley slopes are sufficient to allow the flow to move through a system of one or more head cuts. Bennett (1939) states that there are more than 200 million active gullies in the United States.

The amount of sediment from gully formation, though large, is generally less than that from sheet erosion (Glymph, 1951; Leopold, Emmett, and Myrick, 1966). Some of the gully erosion processes have been described (Ireland, Sharpe, and Eargle, 1939; Brice, 1966), but the cause-and-effect relationships are poorly understood. Thompson (1964), in a study of gully activity at several locations in Minnesota, Iowa, Alabama, Texas, Oklahoma, and Colorado, found an empirical relation in which 77 percent of the variance is explained by four independent variables

$$R = 0.15\,A^{0.49}\,S^{0.14}\,P^{0.74}\,E^{1.00}$$

where R = average annual gully head advance in feet,

A = drainage area in acres,

S = slope of approach channel in percent,

P = annual summation of rainfall from rains of 0.5 inch or more per 24 hours in inches, and

E = clay content of eroding soil profile in percent by weight.

If Thompson's equation is applicable in a given situation, then the amount of sediment moved from an active area would depend on the drainage area, channel slope, and amount of rainfall as factors of energy input, and on the clay content of the eroding profile as a factor resisting the energy input.

Coarse sediment and streamflow

The settling rate, or standard sedimentation diameter, of a particle is a measure of its resistance to transport. In a dispersed state, fine sediment particles are easily carried in complete suspension by the fluid forces in natural streams and hence have a tendency to move out of the drainage basin with the flow in which they are suspended. In contrast, coarse sediment parti-

cles with a relatively fast settling rate may move by suspension for only short distances, or possibly by rolling and bounding along the streambed. The smaller of these coarse particles move with longer step lengths and shorter rest periods, or a faster mean velocity, than do the larger particles with shorter step lengths and longer rest periods. The largest particles in the bed of a given stream would be transported only a short distance in a given period of movement and then only when the stream is experiencing a great flood. The coarse sediments found in abundance on or near a streambed are being continuously sorted by the selective transport capacities of the stream. This selective transport capacity is indicated by the concentration of the different sizes of sediments suspended in the cross section. An example is given in figure 7 for the Missouri River at Kansas City, Mo.

Though the quantity of fine sediment moved by the stream at a given time is nearly equal to that released within the drainage basin, the quantity of the various coarser sizes in transport is closely related to the magnitude of the fluid forces per unit area of the stream channel. For the coarse material, Lane (1955) reported that if the supply is not equal to the carrying capacity through a stream reach, the stream will aggrade or degrade to establish approximate equilibrium between capacity and discharge of coarse sediment within the reach.

Mean velocity and resistance to flow

Sand swept up from the bed of a natural stream or suspended in a stream may be supported by the vertical components of currents in turbulent flow and transported downstream a considerable distance. The magnitude of these currents is largely a function of the horizontal velocity, the bed roughness, and the distance above the streambed. Therefore, the suspended load of sand in a vertical line within a stream cross section can be considered to be a function of the mean velocity of flow.

B. R. Colby (1964a) showed that the discharge of sand in a sandbed stream is closely related to the mean velocity of flow for rivers of a wide range of sizes. Many investigators had previously used the supposedly logical parameter of stage or depth as the independent variable for determining sand transport. The fallacy of the depth-transport concept is that the relation between velocity and depth is poorly defined both for an individual stream and among streams (Dawdy, 1961). Colby (1961) illustrated the complexity of the depth-transport concept by showing that sand transport decreases with increasing depth at a specific low velocity (less than about 1 meter/second) and increases with increasing depth at a specific higher velocity.

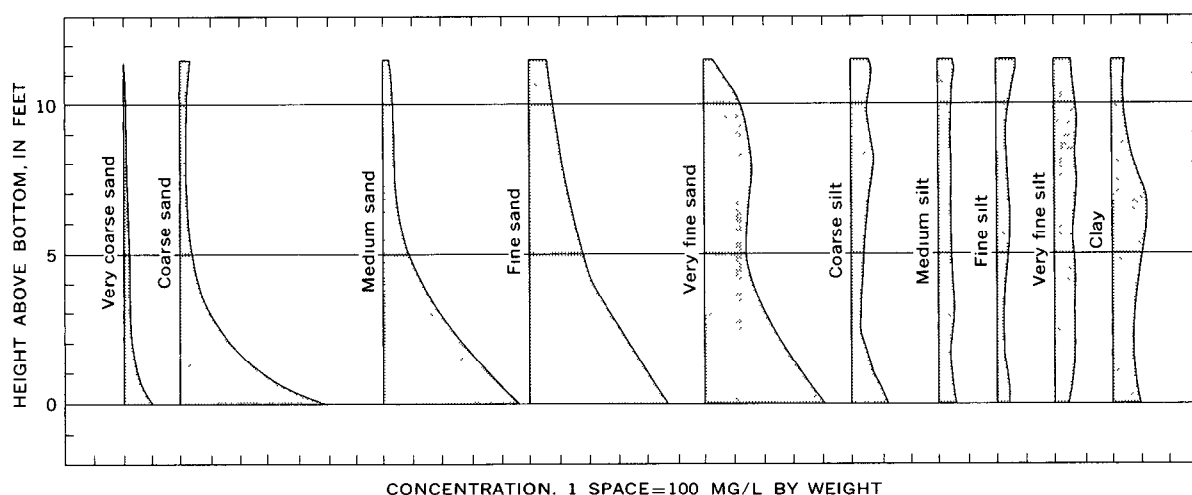CONCENTRATION. 1 SPACE=100 MG/L BY WEIGHT

Figure 7.—Discharge-weighted concentration of suspended sediment for different particle-size groups at a sampling vertical in the Missouri River at Kansas City, Mo.

The complex of transport, depth, mean velocity, and sediment particle size needs to be considered in the light of resistance-to-flow concepts outlined by Simons and Richardson (1962, 1966). They show from flume experiments and observations on natural sand-bed streams that bed forms can be classified on the basis of a lower, a transition, or an upper flow regime. The bed forms that occur are ripples, ripples on dunes, dunes, washed-out dunes, plane or flat bed, antidunes, and chutes and pools. These specific bed forms and the regime classification, as indicated in figure 8, are associated with a specific mode of sediment transport and a specific range of resistance to flow. An example of the effect of bed-material size and Froude number on the form of bed roughness

and Manning n is given in figure 9. In an 8-foot-wide laboratory sand channel, it is noted that ripples generally cause Manning n to range from 0.020 to 0.028; dunes, from 0.020 to 0.033; washed-out dunes, from 0.013 to 0.025; antidunes, from 0.014 to 0.020; and chute and pool, from 0.020 to 0.026 (Guy, Simons, and Richardson, 1966, p. 62–69).

It is important to note that different bed forms and flow regimes may occur side by side in a stream cross section in the form of multiple roughness, or one after another in time in the form of variable roughness. The relatively large resistance to flow in the lower regime results mostly from form roughness whereas most of the resistance in the upper regime results from grain roughness and wave formation and sub-
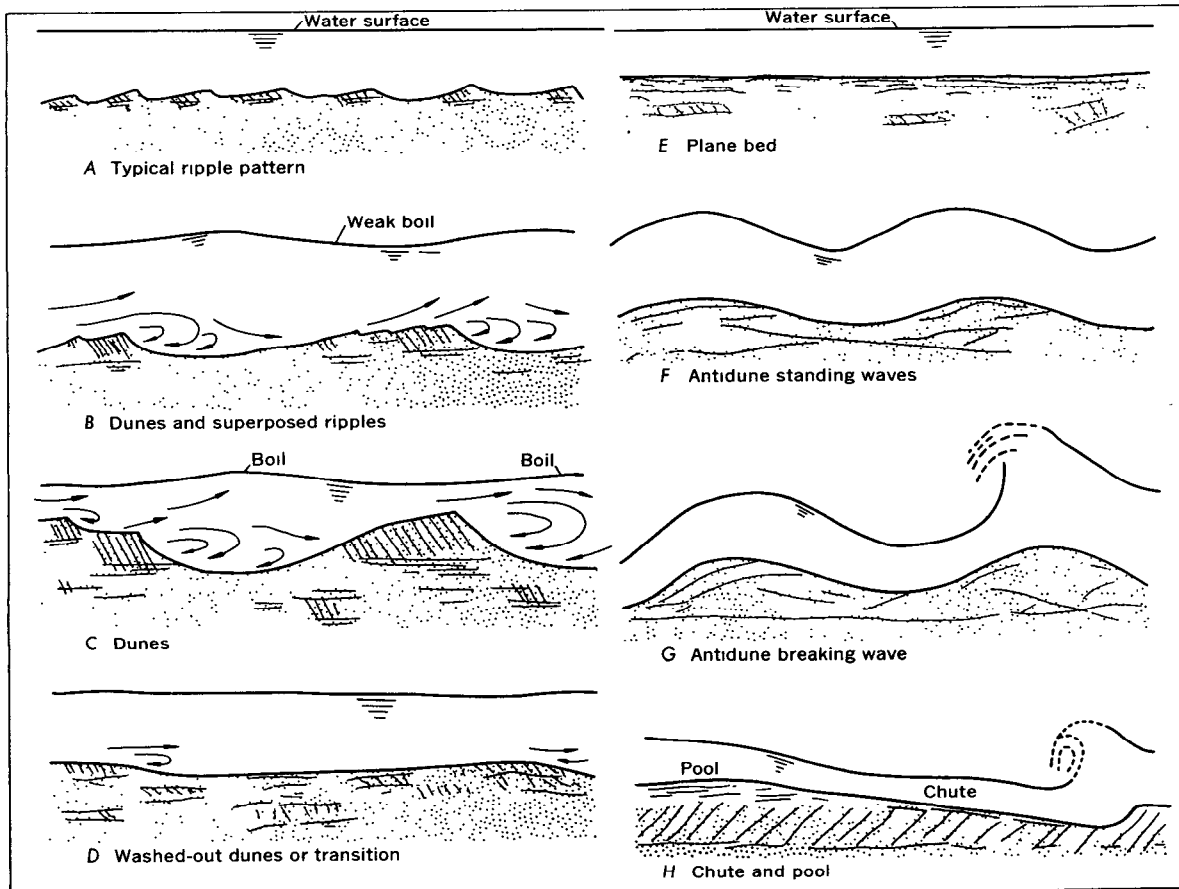
Figure 8.—Schematic diagrams of eight types of roughness found in sand-bed channels. Types A through C are representative of the lower flow regime where the Froude number is usually <0.4, E through H are representative of the upper flow regime where the Froude number is usually >0.7, and D represents the transition regime. Modified from Simons and Richardson (1966, p. J5).
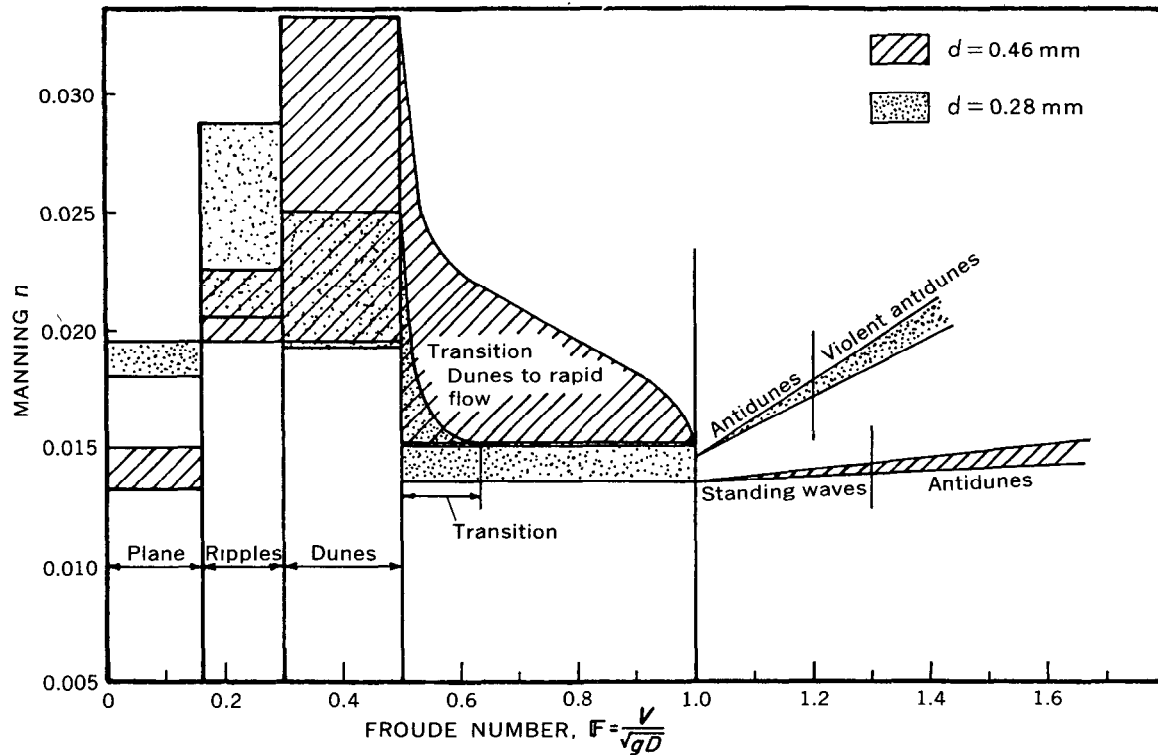
Figure 9.—Effect of size of bed material and Froude number on form of bed roughness and Manning n for a range of flow conditions with sands of 0.28- and 0.46-mm median diameter in an 8-foot-wide flume. Modified from Simons and Richardson (1962, p. E7).

sidence. Resistance to flow for a plane bed is less when the bed material is moving than when the bed material is not moving.

The occurrence of different bed forms in a streambed at a given time and for different times has been discussed by Colby (1964b). This kind of variation is best illustrated by his schematic diagram of bed positions with time at six points in a stream cross section. (See fig. 10.)

Particle movement

In the development of a technique for computation of the amount of sand transport, Einstein (1950) treated the beginning of movement and the pickup of the sand grains from the bed as a probability for the individual grains to move. Thus, a specific critical velocity for "beginning of motion" is probably arbitrary and inexact as a measure of bed movement because of the arrangement of the grains on the

bed and because of local variations of velocity at the bed surface. At a velocity greater than the so-called "critical value," movement in a very thin layer may occur by rolling, sliding, or skipping along the bed.

Equilibrium of the concentration gradient of suspended sediment at a stream vertical requires that particles settling through a horizontal plane be balanced with a net upward movement of particles through this plane from a zone of heavier concentration. Particle fall velocity is then considered to be an indication of the rate of change of sediment concentration with distance above the streambed for a given scale and intensity of turbulence. An increase in turbulence, considered to mean an increase in the vertical movements of flow, causes more uniformity of concentration for a specific size of sediment with respect to distance above the bed. Therefore, high values of turbulence tend toward a uniform vertical concentration of
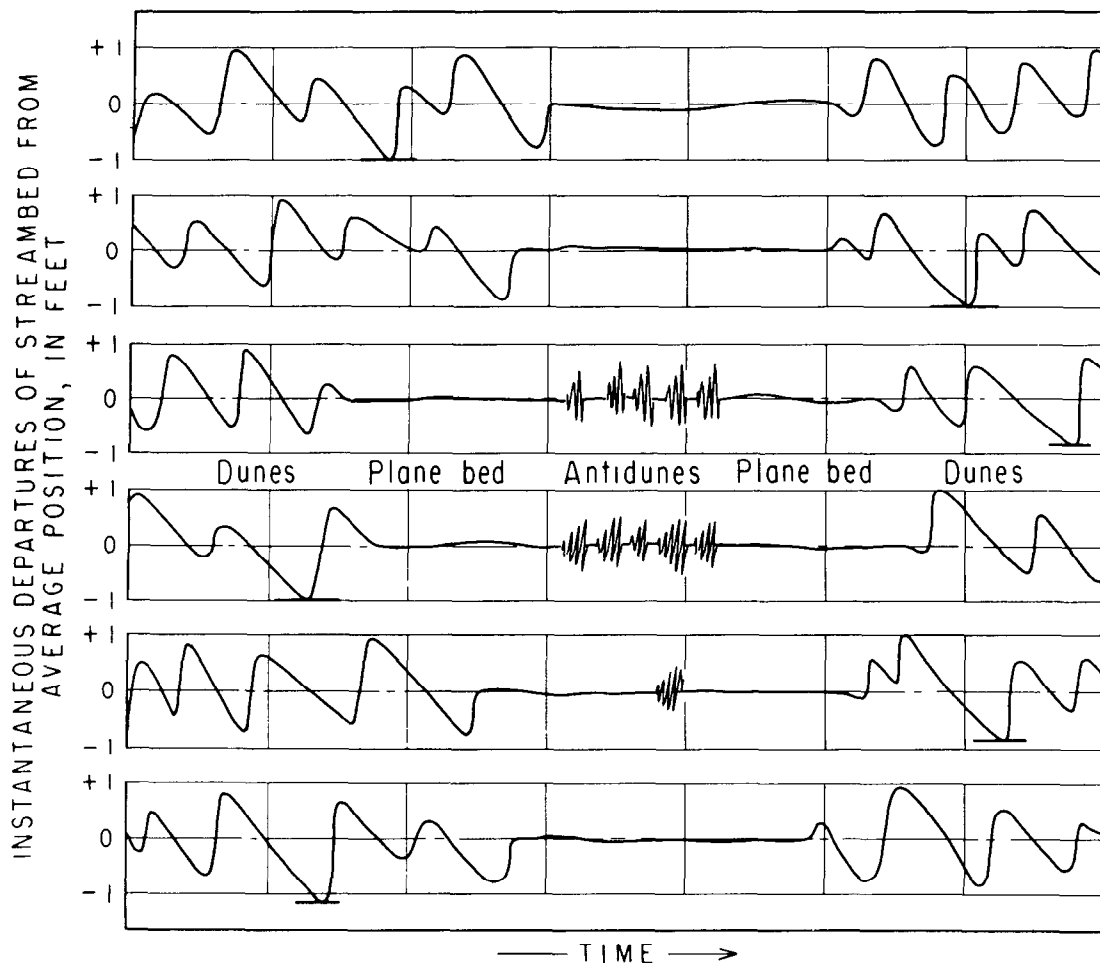
Figure 10.—Schematic diagram of streambed elevation with time at six points in a stream cross section. Time of changes not to scale. From Colby (1964b, p. 4).

sediment. If mean velocity is an indication of the scale and intensity of turbulence and the vertical variation of sediment concentration, then the discharge of coarse sediment is related to both stream velocity and particle size.

Colby (1961) showed that, for a given mean velocity and a given bed roughness, there will be greater turbulence and a higher concentration of suspended coarse particles in a shallow section of a given stream than in a deep section. Averaged over a long period of time, the sediment transported at two separate cross sections of a stream is likely to be equal even though the sections are of dissimilar depth and velocity. With a substantial change in flow characteristics with respect to depth and velocity, the transport through the two sections may temporarily be different, causing aggradation or degradation (fill or scour) of the streambed.

Effect of viscosity

Laboratory studies by Simons, Richardson, and Haushild (1963) show inconclusive results regarding the effect of increasing concentration of fine material on the transport of coarse sediment. However, the data support the conclusion that, for a given bed roughness, an increase in fine-sediment concentration will increase the transport of coarse sediment because the mean velocity of flow may be increased and the fall velocity of sediment particles may be decreased. The change in fall velocity of sedi-

ment particles is caused by changes in the density and "apparent" viscosity of the suspending fluid.

Water temperature is an important environmental factor affecting the transport of coarse sediment, through its effect on viscosity of the fluid and the resulting changes in the fall velocity of the particles and changes in the turbulence of the streamflow. The effect of water temperature change on particle fall velocity is greatest for fine sediment because these sizes settle more nearly in accordance with Stokes' law. For example, particles in a size class of 0.016–0.062 mm have a fall velocity of about 0.051 cm/sec (centimeter per second) at 0°C and 0.116 cm/sec at 32°C, whereas particles in a class of 1.00–2.00 mm have a fall velocity of 1.80 cm/sec at 0°C and 2.26 cm/sec at 32°C (Hubbell and Matejka, 1959). Temperature change, however, does not affect the amount of fine material transported (less than 0.062 mm) because its quantity is limited by the amount supplied to the stream system; that is, the stream will readily carry the entire input of fine sediment at either a high or a low temperature. The temperature effect is probably most important for fine and medium sizes of sand.

Variations in concentration of sediment

As a result of the variations of the rate at which fine sediment moves into streams and the way both the fine and coarse sediment are transported in stream channels, it is evident that a great deal of variation can be expected in the concentration of sediment at a given stream cross section. Such variations can be considered as a function of time at a point within the cross section or with respect to the entire area of the cross section. Concentration can also be expected to vary with location in the stream section at a specific time. To define the stream sediment concentration or rate of sediment transport, it is necessary to understand something of the sediment variation for both area and point conditions at a given stream cross section. This understanding will make it possible to better formulate a measurement program that will yield the desired kinds of sediment data with the desired accuracy.

Concentration definitions

Before further discussion of sediment concentration variations in a stream, it is desirable to recognize several definitions of concentration. Because sediment particles occupy physical space in the stream or any body of water, it is natural to consider concentration in terms of the relative amount of volume occupied. The units for volume concentration might be milliliters per cubic meter, parts per million, or percent. As expected, volume concentration is difficult to measure because of the small size of most sediment particles and the variable way in which sediment deposits consolidate (p. 32).

In the laboratory, the relative amount of sediment in a sample is best determined by weighing. Such weighings include the water-sediment mixture of the sample and the dry sediment after filtration or evaporation. Therefore, a concentration can be determined as the ratio of the weight of dry sediment to the weight of the water-sediment mixture and expressed as a percentage or parts per million by weight. However, to be consistent with units and definitions commonly used for concentrations of other substances, the ratio of dry weight to mixture weight must be converted to a concentration in terms of milligrams per liter or a ratio of dry weight to volume. Because of the space occupied by sediment in a sample of water-sediment mixture, the applicable factor for converting parts per million to milligrams per liter may range from 1.00 at concentrations between 0 and 15,900 ppm to 1.50 for concentrations between 529,000 and 542,000 ppm (Guy, 1969, table 1). These conversion factors are based on the assumptions that the water temperature is between 0 and 29°C, that the specific gravity of the sediment is 2.65, and that the concentration of dissolved solids does not exceed 10,000 mg/l.

If the sample of sediment from a stream is obtained in a manner to give a velocity-weighted concentration, that is, a sample volume proportional to stream velocity, then a sample at a point in the stream should be representative of and proportional to the concentration of sediment in a volume of flow for some area surrounding the point of sampling. Likewise a depth-integrated sample should be pro-

portional to the sediment discharged in some unit width of flow adjacent to the sampling vertical. The velocity- or discharge-weighted sample is possible because the samplers (Guy and Norman, 1970) are designed and calibrated so that the velocity of flow in the intake nozzle closely equals the surrounding stream velocity, and the assumption is made for the depth-integrated sample that the sampler is moved through the sampling vertical from top to bottom and return at a uniform transit rate.

If the depth-integration concept for a single vertical is expanded to several verticals equally spaced all the way across the stream channel and if a uniform vertical transit rate is used at all sampling verticals, it is apparent that the quantity of water and sediment obtained should be proportional to the total streamflow in the measuring section. This technique of making a discharge-weighted sediment measurement is known as the ETR (equal-transit-rate) method. In laboratory flume operations, a discharge-weighted concentration is usually obtained by traversing the nappe of the flow issuing from the flume at a uniform lateral transit rate with a vertical-slot interception device.

The mean discharge-weighted concentration of a stream can be used directly to compute the rate of sediment discharge moving in the stream,

$$Q_s = Q_w \, C_s k$$

where C_s = discharge-weighted mean concentration, in mg/l,

Q_w = stream flow rate, in cubic feet per second, and

k = the conversion factor of 0.0027.

If Q_s is to be expressed in metric tons and Q_w is in cubic meters or metric tons per second, then k is 0.0864.

Another kind of sediment concentration, though seldom used, is computed from a spatial-collection procedure and defined as the relative quantity of sediment contained in an immobilized prism of water-sediment mixture over a specific area of the channel. The chief distinction between velocity-weighted and spatial concentrations is that one is based on sediment and water discharged through a cross section and the other on sediment and water in motion above an area of streambed at a particular instant.

The dissimilarity between spatial and velocity-weighted sediment concentration in a set of flume experiments has been discussed by Guy and Simons (1964). The spatial concentration must be used if the pressure or specific weight of the flow on the streambed is required.

Effect of drainage area

Just as only part of the eroded sediment in a field would be expected to reach a major watercourse, it is expected that the sediment yield of a large basin would be less than the sum of the yields from its subbasins. This generalization may not hold for basins where the lower reaches are degrading as a result of uplift or where there is a lowering of the base level downstream. Aggradation or alluviation is believed to be more common than degradation because of man's effect on increasing erosion. The controlling condition is simply that more sediment is released from the drainage area than the stream system is capable of removing. Also, in basins of more than about 1 square mile, the intensity of precipitation and runoff for a given storm is likely to vary considerably in different parts of the basin, and because erosion and transport increase geometrically with the input variables, it can be expected that the sum of the loads from the subbasins will be greater than would have been obtained from the whole basin reciving an ideal average input.

The effect of drainage area on sediment movement is explained in simple terms by Gottschalk and Jones (1955, p. 138):

Not all of the material eroded in a watershed is moved out. The bulk comes to rest below slopes and on flood plains. It is estimated that less than one-fourth of the materials eroded from the land surface in the united States ever reach the oceans.

The ratio of the amount of sediment carried out of a basin to the gross erosion within the basin is known as the delivery ratio. The delivery ratio of a drainage basin depends on the areial distribution and intensity of runoff, the size and topographic characteristics of the basin including the degree of channelization, and other soil and land use factors, all of which determine the ability of the drainage system to pick up and transport sediment.

For 15 drainage areas ranging in size from 0.61 to 167 square miles in the southeastern piedmont area of the United States, Roehl (1962) found the sediment delivery ratio, Q_{sr}, to be related to the drainage area, A, in square miles; average total stream length, L, in feet; the relief-length ratio, R/L; and the weighted mean bifurcation ratio, BR (page 35), the ratio between numbers of successively higher stream orders. These relationships follow:

$\log Q_{sr} = 1.91 - 0.34 \log 10A$

$\log Q_{sr} = 1.63 - 0.65 \log L$

$\log Q_{sr} = 2.89 - 0.83 \operatorname{colog} R/L$

$\log Q_{sr} = 4.50 - 0.23 \log 10A - 0.51 \operatorname{colog} R/L - 2.78 \log BR.$

The correlation coefficients for these equations are 0.72, 0.81, 0.87, and 0.96, respectively.

The effect of channel aggradation on the downstream diminution of sediment discharge was cited by Borland (1961) for a glacier-fed Alaskan stream. The annual sediment yield for 868 square miles was 9,120 acre-feet or 10.5 acre-ft per sq mi, whereas farther downstream the yield for 6,290 sq mi was 6,440 acre-ft or 1.02 acre-ft per sq mi. The total runoff for the larger area was nearly triple that for the smaller area. Lustig and Busch (1967) report data from 1960–1963 for Cache Creek, Calif., that indicates the suspended-sediment yield at Yolo to be only 64 percent of that at Capay even though the contributing drainage area increases from 524 to 609 sq mi.

Data on the rates of valley aggradation from sediment accumulation are scarce, but in most situations accumulation will range from near zero to as much as 6 feet in 30 or 40 years, as in the instance reported by Schumm and Hadley (1957, p. 170) for the Cheyenne River basin, Wyo., where three different fences were installed across the valley at increasing elevations. A classic record is provided by the Nile which, according to Lyons (1906, p. 315–317), had been building up its bed and flood plain at a rate of about 0.03 foot yer year in the vicinity of Karnak and Memphis. This is about one-sixth the "rapid" rate indicated by the fence-posts in the Cheyenne River basin. Happ, Rittenhouse, and Dobson (1940, p. 21) measured aggradation of 0.12 foot or more per year in small valley bottoms. This aggradation was caused mostly by "accelerated sheet erosion" from agricultural lands.

Leopold, Wolman, and Miller (1964, p. 435) report

The history of hydraulic mining in the Sierra Nevada, Calif., not only illustrates the effect of man on the landforms of a region but also provides a good example of aggradation as a result of increasing sediment yield without compensating increases in flow. In the early days of the gold rush only a small amount of dirt was disturbed, as most of the work was done by laborers with pick and shovel. As more efficient methods were developed, water power was substituted for manpower and vast quantities of earth were handled in separating the gold from the placer deposits in which it was found. Hydraulic mining increased steadily until 1884, when a series of injunctions brought by residents of downstream areas halted the entire operation. At the height of hydraulic mining it is estimated that scores of millions of cubic yards of earth were moved each year. Apart from the considerable topographic changes rendered directly by the mining, the principal effects were those on the streams, which resulted from overloading with detritus and led to extensive aggradation over broad areas.

One cannot estimate the precise effect of aggradation on sediment storage in the basin, but curves provided by Glymph (1951) indicate the trend to be expected. For example, the annual yield from a variety of drainage areas of 5 sq mi (13 square kilometers) generally ranges from 400 to 4,000 tons per sq mi (140 to 1,400 metric tons per sq km), whereas for 500 sq mi (1,300 sq km) the range is 100 to 2,000 tons per sq mi (35 to 700 metric tons per sq km). Furthermore, Glymph cautions,

Too often records of soil loss from plot studies have been erroneously interpreted as a measure of sediment supply with respect to some point of damage lower in the watershed. Similarly, sediment carried by a stream or accumulated in a reservoir has been erroneously interpreted as a measure of erosion in the watershed.

Hydrograph characteristics (time)

Storm or surface runoff is defined as the part of total runoff derived from storm rainfall or rapid snowmelt which reaches a stream channel within a relatively short period of time. The time for such runoff to reach a peak rate at a site depends on many drainage-basin characteristics, the most important of which is probably area. Only a few minutes are required for areas of a few acres, but several days may be required for drainage areas of thousands of square miles.

The groundwater runoff or base-flow part of a streamflow hydrograph lags the causative precipitation by a distinguishably longer period of time than does the surface runoff. Often, storm runoff may include subsurface groundwater flow which has infiltrated the surface of the ground but causes an increase in groundwater flow to the surface channel sufficiently soon to be classed as storm runoff. Such rapid movement of the subsurface storm flow occurs in areas near the stream through perched water tables, through flowing saturated zones, or through semichannels beneath the surface. The true surface runoff, or that amount of precipitation in excess of infiltration and surface storage, reaches a surface channel with its path on and above the ground surface. Except for ephemeral streams and small plots or fields, the delineation of the amount of overland flow is difficult and inexact because there is no way of measuring either the overland flow or the groundwater contribution to the streamflow.

The relationship of the sediment concentration to the hydrograph has been characterized by Colby (1963):

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 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 concentrations.

The variation of concentration with respect to the storm runoff hydrograph may be illustrated by examples showing the advanced, simultaneous, and lagging concentration graphs plotted together with their gage-height graphs. (See fig 11.) It should be emphasized that the advanced type is the most common and that a given drainage basin will usually yield similar graphs for each storm, especially for basins
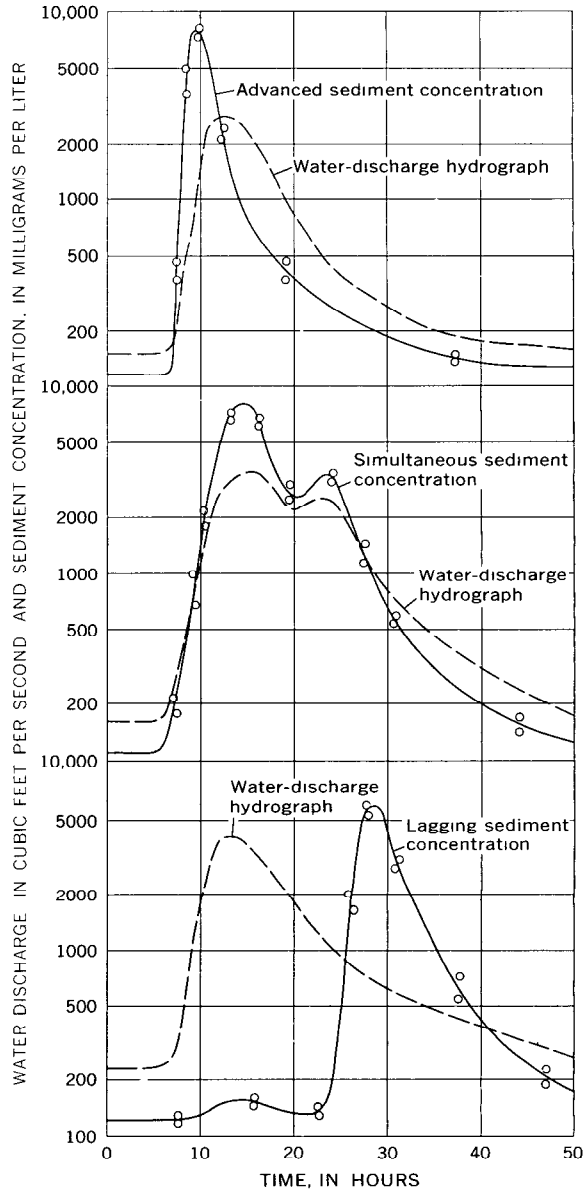

Figure 11.—Advanced, simultaneous, and lagging sediment-concentration graphs as related to the temporal distribution of their respective water-discharge hydrographs. Terms leading, inphase, and delayed are sometimes used.

receiving a relatively uniform precipitation excess. Small drainage basins would not be expected to yield a notably lagging concentration graph. Because of the large change in sediment concentration and the possible change of particle-size distribution during the hydrograph, it is desirable and sometimes necessary

to sample the rising part of the concentration graph on an hour by hour basis (or even minute by minute basis for small drainage areas).

The magnitude of sediment concentration for the "typical" graph at a given stream location will vary considerably depending on the season of the year, the changing patterns of land use, the antecedent moisture conditions, and the nature of the precipitation intensity and pattern on the basin. The concentration graph will also vary a great deal among different drainage basins because of differences in climate, geology, and land use. The potentional seasonal change in stream sediment concentration in terms of the erosion index for different locations along the Atlantic coast of the United States is illustrated by figure 12. The seasonal change in sediment yield would be expected to be different depending on the seasonal variation in the amount of runoff.

It has been mentioned above that sediment yield generally increases geometrically with storm runoff rate. Because storm runoff rate and storm quantity tend to be related, the question arises as to the relative role of the larger storms in contributing sediment from a drainage basin. In a study of 72 small basins in 17 states, Piest (1965) found that large storms (with a return period of 1 year or more) contributed an average of 31 percent of the total sediment yield from their respective basins. The large-storm yield for all basins had a standard deviation of 13 percent within an absolute range of 8 to 66 percent.

For streams in semiarid regions that receive most of their runoff from annual snowmelt, the storm hydrograph may be rather insignificant. The annual hydrograph for a snowmelt type of stream is indicated in figure 13. For this kind of stream, the sampling program can be changed from day to day to coincide with temperature or rate of melting during the early part of the period, usually beginning in March or April. The first few increasing-flow days in
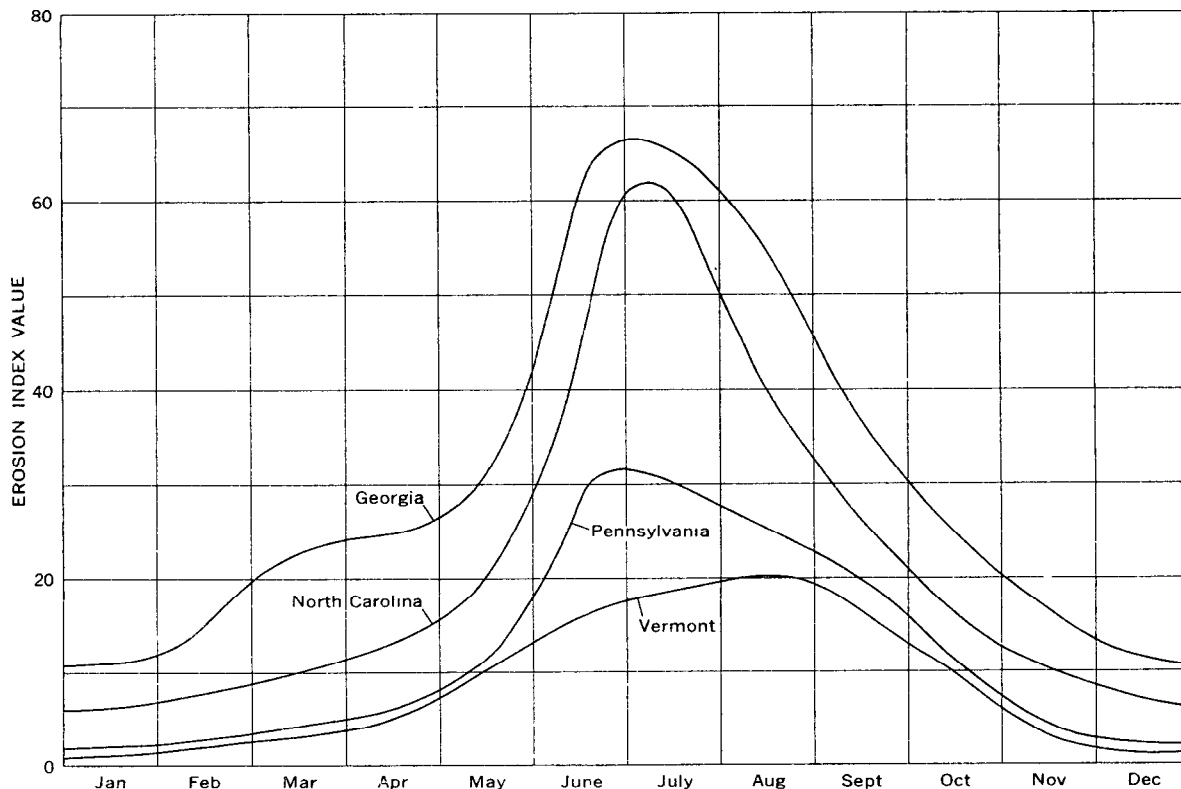

Figure 12.—Seasonal distribution of Wischmeier's erosion-index values at four locations in the Atlantic coast area. From Guy (1964, p. 10).
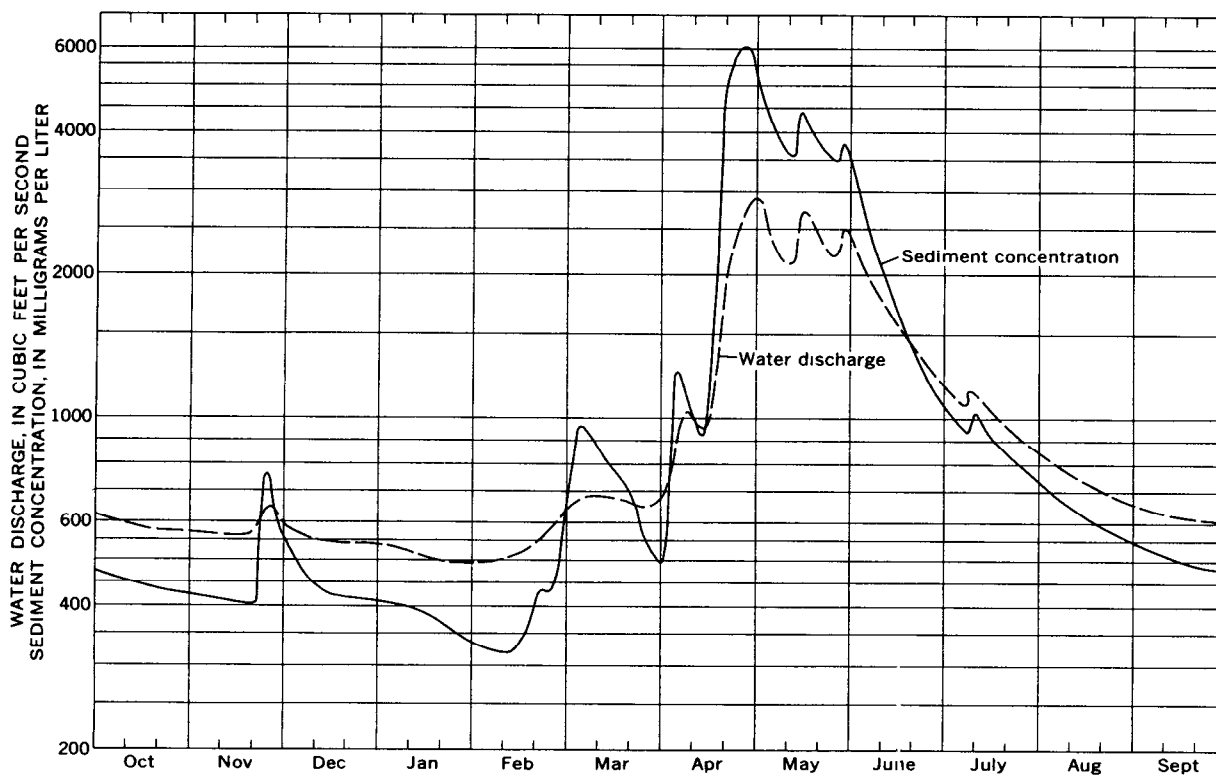
Figure 13.—Temporal relationship of sediment concentration to water discharge for an assumed "snowmelt" stream draining mountainous terrain.

the spring should receive special attention because the stream will likely contain considerable fine sediment loosened by freezing and thawing and mass wasting. The last part of the melt during the summer is expected to transport mostly sand-sized material. During the relatively dry period beginning in September or October, daily samples are not necessary and therefore samples sufficient to define the diurnal fluctuations on perhaps 2 days per month may be adequate.

Cross-section variations

As mentioned, fine sediment is easily suspended by the forces of streamflow and is, therefore, dispersed throughout the stream cross section according to the laws of suspension dispersion (Yotsukura, 1968). For most streams, the mixing length required downstream from a confluence would be roughly the ratio of the mean velocity times the square of the required mixing width to the mean depth of flow. In many instances, however,

complete mixing may not be necessary either because the sediment contribution from the side tributary is relatively small or because the sampling program designed for the coarse sediment will result in an adequate sampling program to define the fine-sediment differences in the cross section.

Coarse sediment, on the other hand, is not easily or completely suspended by streamflow and therefore, at a specific location in the stream cross section, moves in accordance with the hydraulic characteristics of the flow. As mentioned on pages 15 and 16, sand transport or suspended-sand concentration variation needs to be considered in the light of resistance-to-flow concepts. This means that the flow regime and bed forms are important (fig. 8). The maximum lateral, vertical, and temporal variation in sand suspension can be expected over a dune bed, whereas the minimum variation can be expected over a plane bed. As already stated, the problem is complicated by the fact that considerable variation

of the specific bed form or roughness is likely to occur across the section and with time at a given location.

It is then evident that coarse-sediment movement through a stream section is difficult to define because of the variation at a vertical over the bed with time as well as the variation across the section at a given time. The measured variation with time for 20 consecutive samples collected at each of two separate verticals on a dune bed and on a plane bed of the Middle Loup River at Dunning, Nebr., is illustrated in figure 14. The relative sand-concentration variation at most streams would be expected to range between these two examples.

Assuming that the mean concentration of coarse sediment at each of several verticals across the stream can be measured, it is then possible to determine the nature of the lateral concentration variation. As expected, the greatest variation occurs with the roughest dune-bed condition. Measurements of the Middle Loup River at Dunning, Nebr., show the lateral distribution for two sets of samples taken only a few minutes apart on each of two occasions about 6 weeks apart. (See fig. 15.) The lateral distribution of the water discharge is indicated for the samples only on November 24, 1955, because the water-dis-

charge data were not obtained at the time of sampling on January 7, 1956. The data presented in figure 15 may not be representative of the roughest dunes and shallowest depths but are likely to be typical of many sand-bed streams.

If a sand-bed stream typically moves large quantities of fine sediment in addition to the coarse during high-flow periods, the variation of total concentration will be much less with respect to both consecutive and lateral samples than for the condition of mostly sand transport. For example, the overall sediment-concentration variation would be reduced to as little as one-fourth the normal coarse-sediment variation if the fine-sediment concentration were increased to four times the coarse-sediment concentration, assuming, of course, that the fine sediment were dispersed uniformly in the cross section.

In this discussion of sediment-concentration variations in the cross section of a sand-bed stream, the assumption is made that the concentration will be defined by depth-integration techniques whereby the sample intake is proportional to the stream velocity at all times. Again, if only fine sediment were involved, this assumption would generally not be important; but for coarse sediment, the concentration from the water surface to a point 0.3 foot (10 cm)
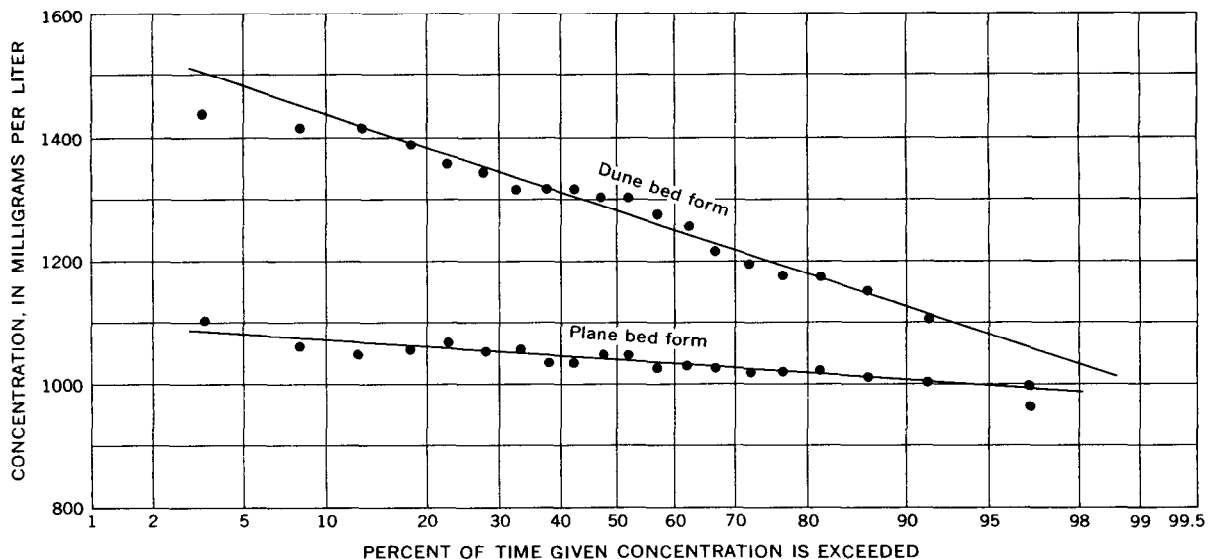

Figure 14.—Frequency distributions of consecutive sampled concentrations at single verticals of the Middle Loup River at Dunning, Nebr.
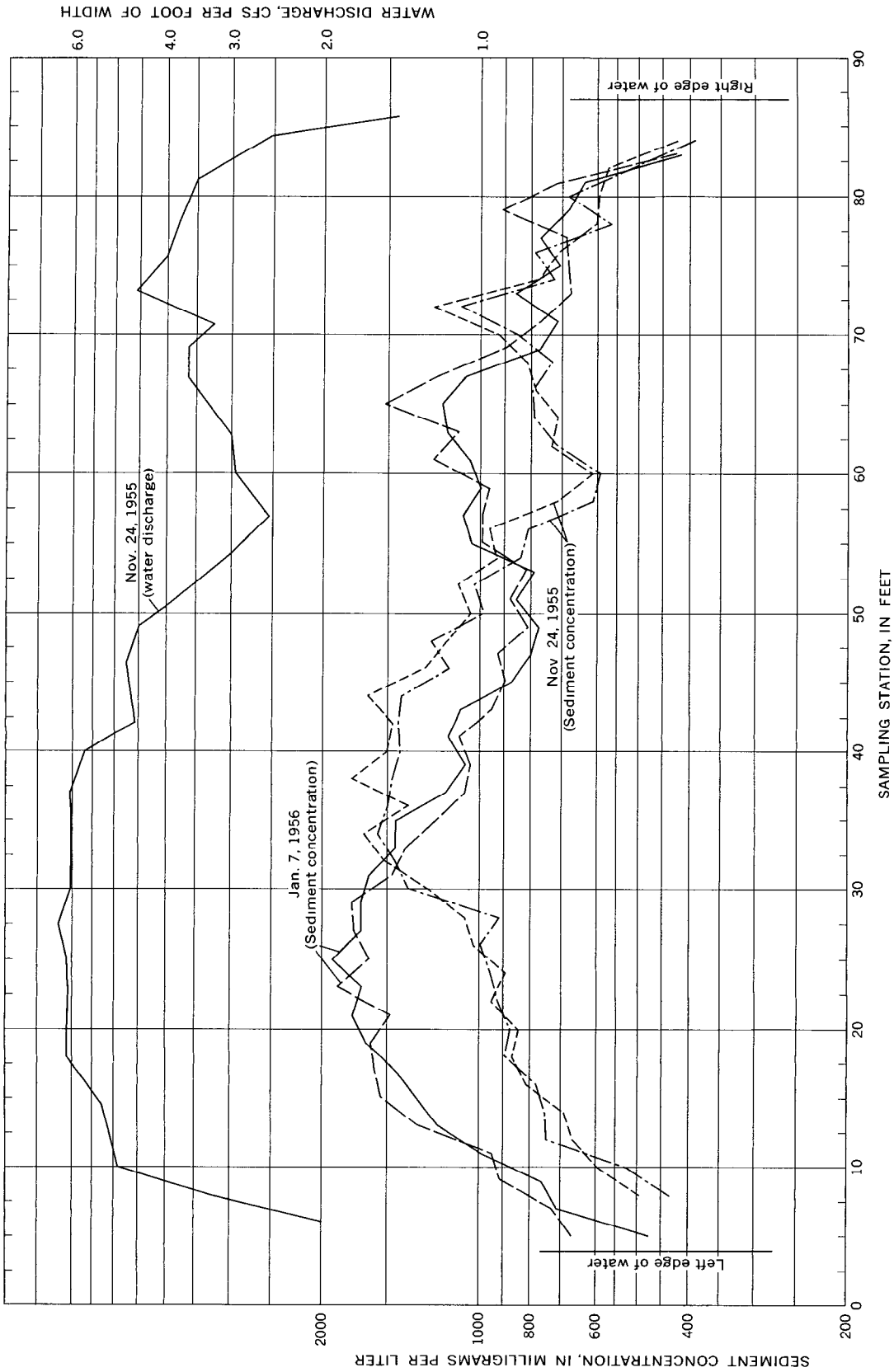
Figure 15.—Lateral variation of sampled sediment concentration for two different days on the sand-bed stream, Middle Loup River at Dunning, Nebr. The two measurements on each day were taken only a few minutes apart. Modified from Hubbell (1960).

above the bed may possibly range from 0 to over 106,000 mg/l. The concentration at a given level will depend largely on the stream depth and velocity characteristics, the bed form, and the sediment characteristics. If it is necessary to define the concentration distribution in the sampling vertical, it must also be recognized that considerable variation from second to second will occur at a given sampling point and therefore, to define a representative mean concentration at the point, the 20- to 40-second or longer sampling time may be needed.

Deposition

As implied in the discussion of sediment particle size (p. 9), sediment deposition depends on the particle fall velocity and the dynamic hydraulic characteristics of the suspending medium. In still water, as in a reservoir, the depositional rate of sediment particles may be nearly the same as the fall velocity measured in the laboratory whereas in turbulent stream-flow, the same particles will be dispersed upwards as well as downward even though the net downward movement may be nearly the same as that for still water.

The following from Colby (1963, p. 32) will dispel any notions that a stream will rapidly clear itself of sediment because of the net downward movement of sediment particles:

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 may quickly cause net disposition.

More specifically, the vertical motion of suspended sediment between two levels in a stream may be described in terms whereby a volume of mixture from an upper level having a given concentration is exchanged with an equal volume from a lower level having a greater concentration. This kind of continuous exchange between zones of lesser concentration above and greater concentration below is in an equilibrium or balance with the constant fall velocity of the sediment that occurs while the exchange of mixture is occurring between the two levels. Thus, in flow with much turbulence and (or) particles with a low fall velocity, the concentration gradient between levels would be small, whereas in flow with little turbulence and (or) particles with a high fall velocity, the concentration gradient would be large. This concept may be complicated somewhat where particles are close enough together (high concentration) to interfere with isolated motion or where the chemical quality of the water may cause flocculation of clay particles.

Location of deposits

Sediment deposition may occur at any point in the flow system from (1) sources very near the point of erosion, as in a cultivated field, at the base of a cut slope along a highway, at a road drainage culvert, and across a roadway on which eroded material was deposited from adjacent burned-over foothills (fig. 16 A, B, C, and D), to (2) deposits in stream channels as illustrated in pictures from Scott Run, Va., Montlimar Creek, Ala., Mill Creek, Calif., and the Mississippi River (fig. 17 A, B, C, and D), and to (3) deltaic deposits in larger bodies of water as in the Mississippi River in Iowa, a farm pond in Virginia, Lake Pillsbury, Calif., and Seaman Reservoir, Colo. (fig. 18 A, B, C, and D).

As a result of man's activity in the form of highway maintenance and the cultivation of fields, deposits of the kind shown in figure 16 are likely to be noticed for only a few days or months. Channel deposits generally have a relatively short life because they can be eroded by streamflow from the side of the deposit as in figure 17B and D or from the upper surface during another stage of flow. Unlike the deposits illustrated in figures 16 and 17, deposits in lakes and reservoirs below the lowest operating
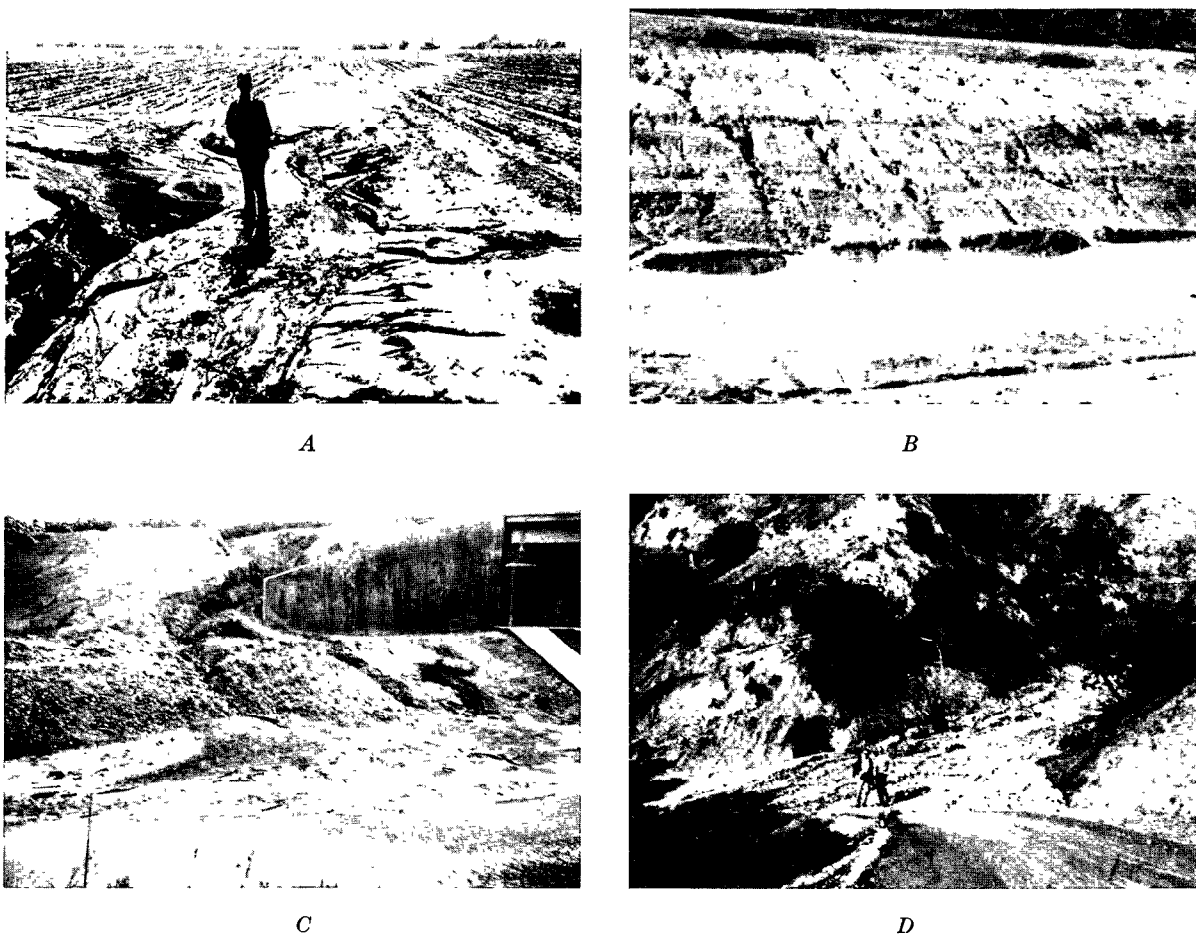
Figure 16.—Examples of sediment deposition very near the source of erosion. A, Erosion and deposition in cultivated field B, Rill erosion on and deposition at the base of a cut slope for a highway near Fairfax, Va. C, Sediment deposition in a channel at a road drainage culvert. D, Sediment deposition across a roadway on which eroded material was deposited from adjacent burned-over hills near Los Angeles, Calif.

level are seldom disturbed, either by man or nature, unless the dam breaks or the sediment must be removed to conserve space for the storage or movement of water.

Because of the sorting processes during erosion, transport, and deposition, it is easy to understand why specific sediment deposits are composed of a unique assortment of particle sizes. Sorting may be rather poor in a deposit at the base of a highway cut slope where the slopes are large and the concentration of sediment in the flow is very high; on the other hand, the sorting may be very good for deposition in a reservoir from inflowing river sediments. As expected, the deposits within the channels of most streams are sorted to only a slight degree and generally for a short time because of the rather changeable spatial and temporal flow patterns of the stream. The more extensive nature of larger streams and their more long-lasting flow patterns will generally result in more extensive and intensive sorting than can be expected in smaller streams. Likewise, on a given stream, a large flood will generally result in more extensive sorting and long-lasting deposits than can be expected for a small flood. Some deposits buried deeply in a bar on a convex bank of a stream or deposited on a flood plain during the period of intensive flooding may remain undisturbed for many centuries.
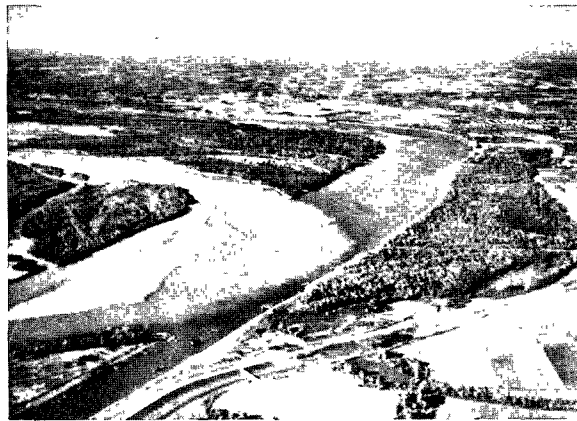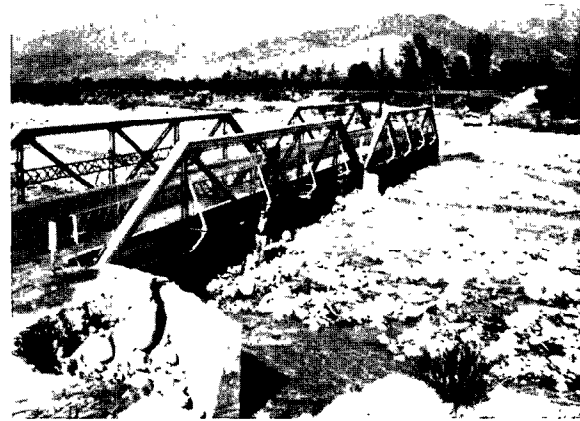
Figure 17.—Examples of sediment deposition in stream channels: A, Scott Run near Fairfax, Va. B, Montlimar Creek at Mobile, Ala. C, Mill Creek near Montrose, Calif. D, Mississippi River at confluence of Missouri River (photograph, courtesy of Massic, Missouri Resources Division).

Reservoir deposition

Though the many kinds of stream sediment deposits may, locally or in aggregate, be of considerable importance, most of the attention has been given to deposition in lakes and reservoirs. Brown (1948) has estimated that loss of storage in reservoirs used for power, water supply, irrigation, flood control, navigation, recreation, and other purposes costs about $50 million annually in the United States. This estimate is based on the value of dollars in 1948 and on surveys of 600 of the 10,000 reservoirs existing at that time. It is also worth noting that reservoir loss measured relative to the initial cost of the structure is not the true economic cost to society because the reservoir is usually constructed at the most favorable site, and therefore, a replacement would be more costly than the original, if at all possible.

Because of the rather extensive study of reservoir deposits, it is possible to glean from the literature some useful concepts regarding such deposition. This information includes such studies as K-79 Reservoir, Kiowa Creek basin, Colo. (Mundorff, 1968), Lake Mead, Ariz. (Smith and others, 1960), and many other reservoirs (Spraberry, 1964). The rate of depositional filling of the reservoir may range from complete filling in a single storm event to negligible filling in several decades. In the example

of K–79 Reservoir, a storm on July 30, 1957, caused deposition of 2.4 acre-ft of sediment; at that time the trap efficiency of the reservoir was about 60 percent. Deposition from this storm occupied about 2 percent of the total reservoir capacity. Mundorff also notes that for such small reservoirs, storms of smaller magnitude have a higher trap efficiency; that is, a smaller percentage of the inflowing sediment is discharged through the spillway. In the example of Lake Mead, 1,438,000 acre-ft of sediment was deposited below the level of the permanent spillway between 1935 and 1948 for a total reduction of 3.2 percent in water storage capacity in a 14-year period. Though turbidity currents carry considerable fines through the reservoir toward the dam, the trap efficiency of Lake Mead, as for other large reservoirs, is very near

100 percent. In the design of small reservoirs, Geiger (1965) reports that the U.S. Soil Conservation Service uses curves developed by Brune (1953) that relate the percentage of sediment trapped to the capacity-annual inflow ratio of the reservoir. The median curve ranges from 45 percent at a ratio of 0.01 to 97 percent at a ratio of 1.0. In design practice, the curve is adjusted upward for highly flocculated and coarse sediments and downward for colloidal and dispersed fine sediments.

The general aspects of reservoir deposition have been describe by Porterfield and Dunnam (1964, p. 9) as follows:

Reservoir sedimentation is a complex process dependent on many factors, and the interaction of the factors may make the sedimentation of each reservoir a case unto itself. The quantity of suspended sediment
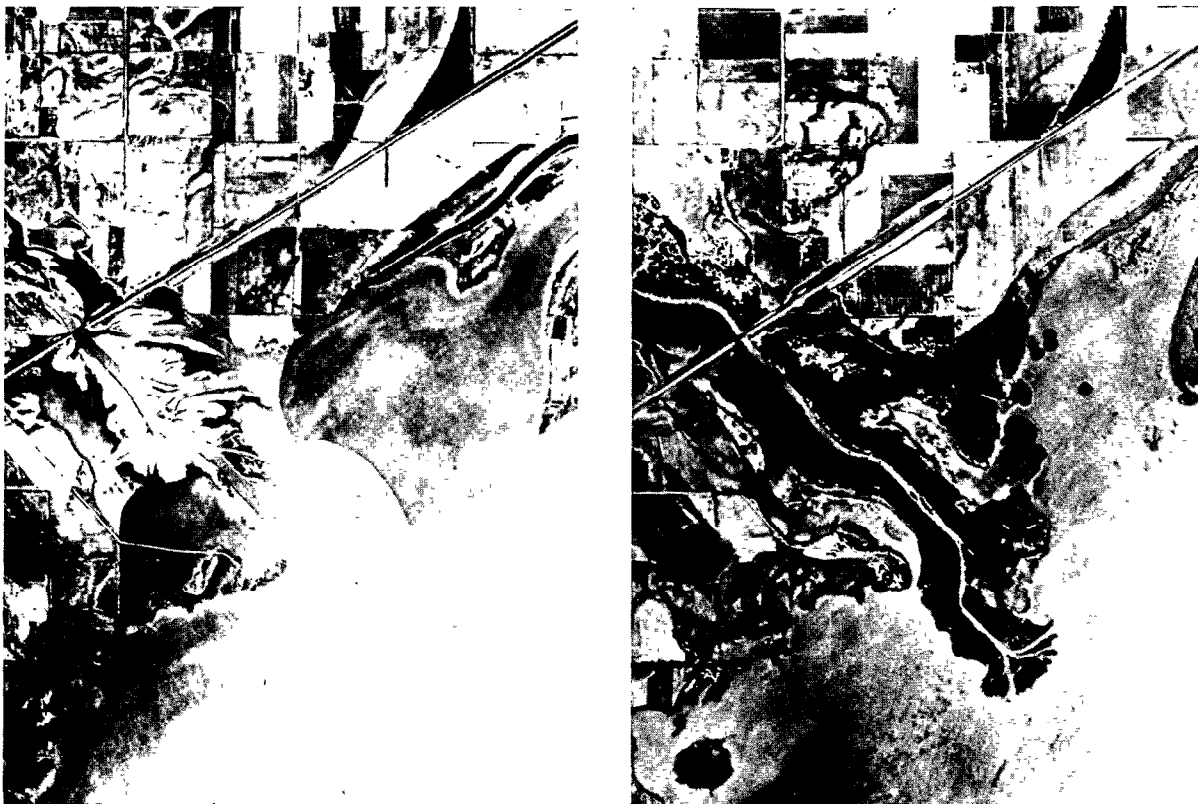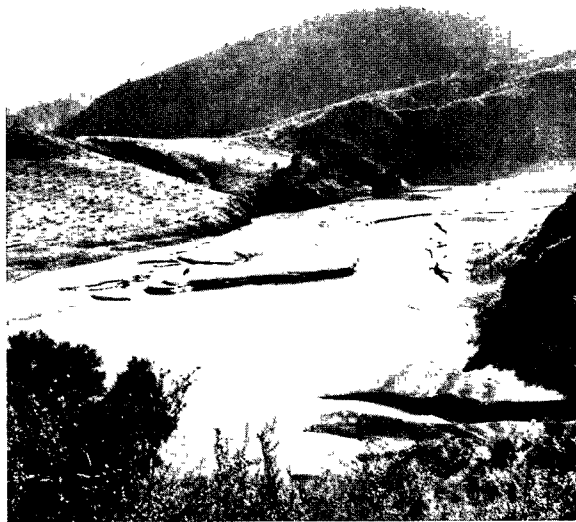

A

Figure 18.—Examples of sediment deposition in deltas: A, Mississippi River at mouth of Devil's Creek, Lee County, Iowa; (left) 1930 conditions, (right) 1956 conditions. B, Farm pond, Fairfax County, Va. C, Lake Pillsbury (Eel River arm), Calif. (photograph, courtesy of George Porterfield). D, Seaman Reservoir on North Fork Poudre River, Colo.

B

C

D

and bed material that moves down a stream can be determined, in most cases, with a fair degree of accuracy, and this knowledge should be utilized prior to the design and construction of any reservoir. However, reservoir sedimentation rates computed strictly from volume of sediment entering the reservoir may be in error (Lane, 1953) because some of the material may flow through the reservoir without deposition and some of the deposition may take place above the spillway elevation of the reservoir. The origin, transportation, and deposition of sediment in reservoirs is discussed by Witzig (1943).

The distribution of the sediment, in addition to volume of sediment deposited, may shorten the life of, or damage, a reservoir. The factors commonly associated with the distribution of sediment in a reservoir are reservoir operation, reservoir shape, wave-action deposits, capacity of the reservoir in relation to amount of inflow, density currents, and properties of the sediment. Additional factors associated with distribution of sediment in a particular reservoir are narrow necks within the reservoir area, vegetation in the delta areas, heavy sediment-contributing streams entering the reservoir area, and the water-surface elevation at the time of maximum sediment inflow.

How sediment is deposited in reservoirs is illustrated in figure 19 (Lane, 1953). The bottomset beds are composed of fine material that is carried into the lake in suspension and settles slowly and somewhat uniformly over the bottom. The density currents, or gravity flow, will move some of the fine material along the bottom far into the reservoir and will produce an additional accumulation near the dam. The foreset beds are composed of coarser material and are inclined downward in the direction of flow. Generally, the angle of inclination of the foreset beds is greater with very coarse sediment than with moderately coarse sediment. The topset beds are composed of the coarsest sediments and extend from the point in the stream where the backwater effect of the lake becomes negligible to the edge of the foreset beds.

Sediment deposits in lakes and reservoirs can quantitively be expressed in terms of either volume or weight. If volume is used, as it is for most deposits, both the solid constituents and the interstitial water or gas must be considered. If weight is used, as it is for most stream-transported sediments, then only the weight of the solid particle is included. For a given set of deposition conditions and a given kind and size of sediment, a relationship between mean specific weight and particle diameter can be developed. Mundorff (1966, p. 31), in a study of deposits in reservoirs for Brownell Creek Sub-
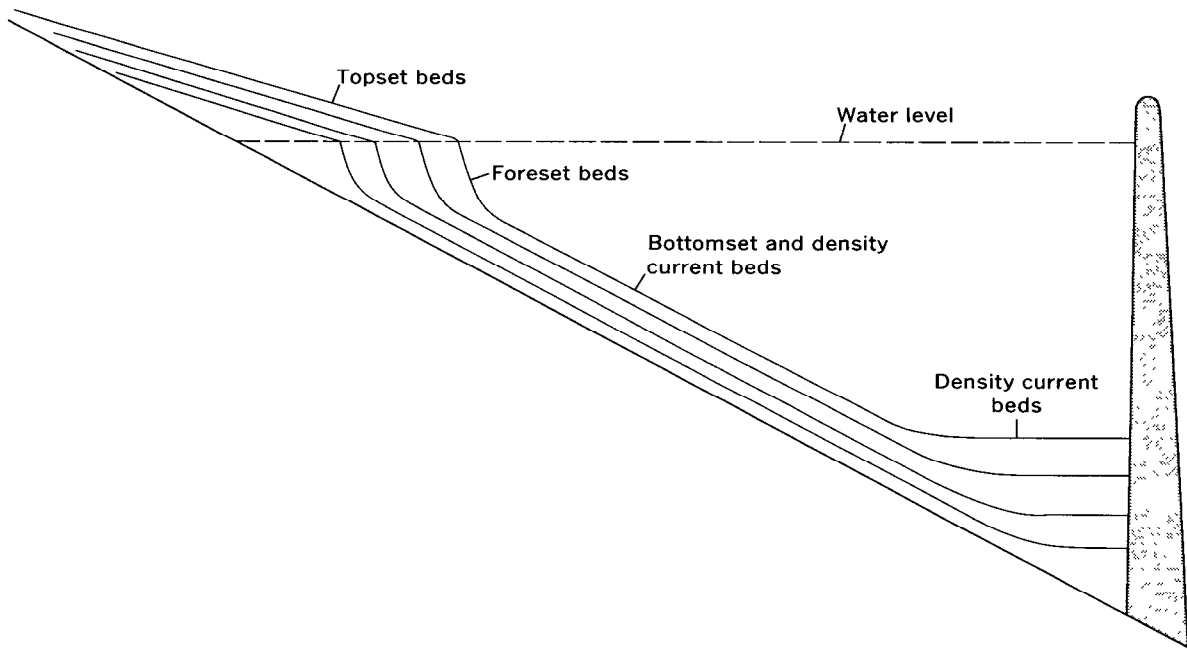
Figure 19.—Longitudinal cross section through a reservoir operating at constant water level. Various types of deposits are shown. Modified from Lane (1953).

watershed No. 1, Nebr., related bulk density in grams per cubic centimeter to the percentage of sand in the sample. In a plot of 18 observations, he found that the higher bulk densities had the higher percentages of sand and that the lower bulk densities had the lower percentages of sand, although there was considerable scatter. Table 3 lists the mean specific weight and median diameter of particles from the different areas in Lake Mead (Smith and others, 1960, p. 196). Based on the volume of sediment repre-

Table 3.—Mean specific weight and median particle diameter for sediments from individual basins of Lake Mead

Area	Mean specific weight, in lb per cu ft	Median particle diameter, in microns
Boulder Basin	34. 1	0. 95
Virgin Basin	39. 8	1. 25
Temple Bar area and Virgin Canyon	41. 8	1. 40
Gregg Basin	49. 4	2. 45
Grand Bay	52. 3	6. 60
Pierce Basin	68. 2	25. 0
Lower Granite Gorge	94. 5	150. 0
Overton Arm (Virgin delta)	78. 2	49. 0
Total basin weighted average	64. 9	46. 0

sented by each of these sizes and weights, the average specific weight of all the sediment accumulated in Lake Mead is 65 lb per cu ft (pounds per cubic foot), and the sediment has a median size of 0.046 mm. The mean specific weight of the sand in the topset and foreset beds is 94 lb per cu ft, and the mean specific weight of the silt and clay in the bottomset beds is 52 lb per cu ft. The sediment in the Virgin delta averages 78 lb per cu ft, whereas the material in the Colorado delta averages only 65 lb per cu ft.

From both field and laboratory studies (U.S. Inter-Agency Report, 1943), it is evident that the specific weight of a sediment deposit will be affected by the size and gradation, by time (especially for fine sediment), and perhaps by the environment in which the deposits are formed. Figure 20 shows the relationship of specific weight to particle size for several different studies of deposits either from different environments or at different times of settling or in which different measures of particle size were used. For a given pressure, drying or aeration of the deposit helps to accelerate consolidation through removal of the water from the pores
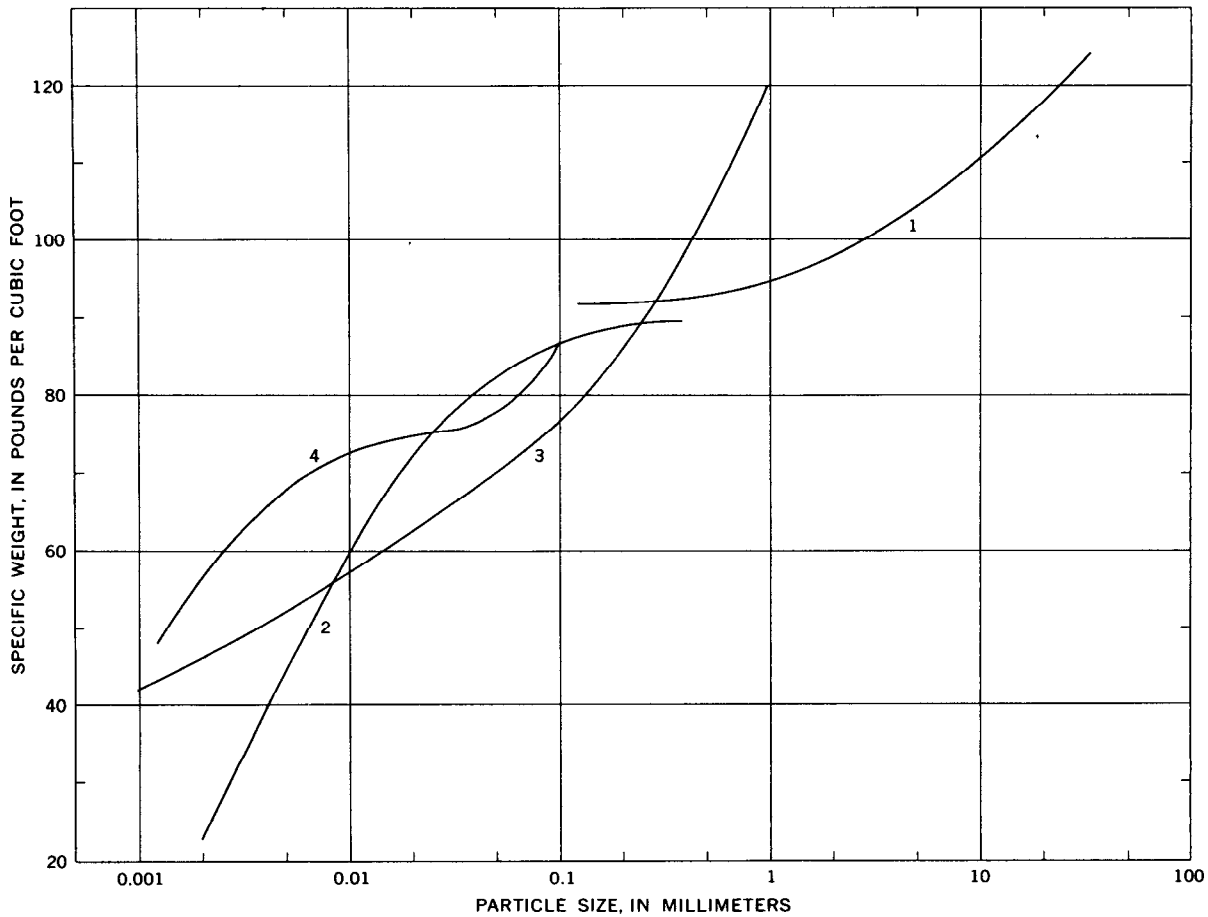
1. California (Koelzer and Lara, 1958)   3. Countrywide (Hembree and others, 1952)
2. Laboratory studies by Parker Trask (U.S. Inter-Agency Report 9, 1943, p. 30)   4. Middle Rio Grande (Happ, 1944)

Figure 20.—Specific weight of sediments of various sizes that have been deposits for 1 year or less. Relation 1 is based on the maximum 10 percent size, 2 is based on specific size classes, and 3 and 4 are based on the median size.

between the grains. Table 4 as published by Geiger (1965) shows the effect of aeration on the specific weight of reservoir sediments for several dominant size classes.

It is also important to recognize that the sediment capacity of a reservoir is greater than the water capacity because sediment deposition will slope upstream from the location of the coarse-fraction delta deposits at a slope somewhat less than the slope of the original stream channel. In other words, the deposition delta will increase in height and extend upstream as the delta or foreset beds proceed through the reservoir toward the dam. Such a delta may be severely eroded by inflowing water and sediment

if the water level is lowered considerably below the level for which the delta was formed.

Table 4.—Ranges in weight-to-volume ratio of permanently submerged and aerated reservoir sediments of specific size classes

[Figures given in pounds per cubic foot]

Dominant grain size	Permanently submerged	Aerated
Clay	40 to 60	60 to 80
Silt	55 to 75	75 to 85
Clay-silt mixture	40 to 65	65 to 85
Sand-silt mixture	75 to 95	95 to 110
Clay-silt-sand mixture	50 to 80	80 to 100
Sand	85 to 100	85 to 100
Gravel	85 to 125	85 to 125
Poorly sorted sand and gravel	95 to 130	95 to 130

Denudation

The net result of sediment erosion, transport, and deposition is a leveling of the continents, because all transport is toward a lower level. Though denudation rates are highly variable over a given area, they are generally expressed as a uniform lowering of the land surface in feet or inches per 1000 years, or years per foot. Usually, the dissolved-solids load of a stream accounts for a considerable part of denudation. The dissolved-solids and sediment yield of stream basins is usually measured in terms of tons per square mile per year. Therefore, using a minor rearrangement of an equation presented by Dole and Stabler (1909),

$$D = 0.0052 \ Q_s,$$

where D is denudation in inches per 1000 years and Q_s is sediment yield in tons per square mile per year.

Rates of denudation, based on both dissolved-solid and sediment loads for seven regional areas, are given in table 5 as previously published by Judson and Ritter (1964). These areas include all the United States except the drainage of the Great Basin, the St. Lawrence River, and the Hudson Bay areas. Holeman (1968) has used this information together with other fluvial-sediment data around the world to show that about 20 billion tons of sediment is transported to the oceans each year. This represents 2.7 inches per 1000 years of denudation and an average yield of 520 tons per sq mi. The Holeman estimate is close to Schumm's (1963) estimate of 575 tons per sq mi and 3 inches per 1000 years.

Geomorphic aspects

Rains occur even in the most absolute deserts, though infrequently. Thornbury (1954) suggests that even desert landforms are mostly the work of running water. Some understanding of the geomorphic aspects of drainage areas will assist in the work of obtaining useful fluvial sediment data. Likewise, as indicated later, good fluvial sediment data will be useful to the geomorphologist.

The drainage basin

The drainage basin forms the natural unit for geomorphic consideration with respect to fluvial sediment. Drainage of excess rainfall from the basin occurs as overland or sheet flow by gravity across the planelike upland areas; with sufficient accumulation of depth and velocity, erosion occurs to form a network of drainage channels. The detail and extent of the recorded drainage system frequently depends on the detail of the map used. The network may be described in various venation terms such as trellis or palmate.

Small rills are integrated into a drainage net on a fresh surface by cross grading and micropiracy (Leopold and others, 1964, p. 411). Cross grading occurs during very heavy storms when water overtops the rill divides and erodes paths that reduce the flow in the upper rill and increase the flow to an adjacent lower rill. Micropiracy may occur with smaller storms when a small channel's drainage system is robbed by a larger channel. Further development of the drainage net will take place as each new com-

Table 5.—Regional denudation in the United States

Drainage region	Drainage area (1,000 sq mi)	Average load (tons per sq mi per year)		Total denudation (inches per 1,000 years)
		Dissolved solids	Sediment	
Colorado River	246	65	1, 190	6. 5
Pacific slopes	117	103	597	3. 6
Western Gulf of Mexico	320	118	288	2. 1
Mississippi River	1, 250	110	268	2. 0
South Atlantic and eastern Gulf of Mexico	284	175	139	1. 6
North Atlantic	148	163	198	1. 9
Columbia	262	163	125	1. 5