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

UNITED STATES DEPARTMENT OF THE INTERIOR

CECIL D. ANDRUS, Secretary

GEOLOGICAL SURVEY

W. A. Radlinski, *Acting Director*

First printing 1970

Second printing 1973

Third printing 1978

UNITED STATES GOVERNMENT PRINTING OFFICE : 1970

For sale by the Branch of Distribution, U.S. Geological Survey,
1200 South Eads Street, Arlington, VA 22202

PREFACE

The series of manuals on techniques describes procedures for planning and executing specialized work in water-resources investigations. The material is grouped under major subject headings called books and further subdivided into sections and chapters; Section C of Book 3 is on sediment and erosion techniques.

The unit of publication, the chapter, is limited to a narrow field of subject matter. This format permits flexibility in revision and publication as the need arises.

Provisional drafts of chapters are distributed to field offices of the U.S. Geological Survey for their use. These drafts are subject to revision because of experience in use or because of advancement in knowledge, techniques, or equipment. After the technique described in a chapter is sufficiently developed, the chapter is published and is sold by the U.S. Geological Survey, 1200 South Eads Street, Arlington, VA 22202 (authorized agent of Superintendent of Documents, Government Printing Office).

CONTENTS

FIGURES

TABLES

ENGLISH-METRIC CONVERSION TABLE

[For fluvial sediment measurements]

Length:

Inches × 0.0254 = meters; . . . × 2.54 = centimeters; . . . × 25.4 = millimeters.
Feet × 0.3048 = meters; . . . × 30.48 = centimeters.
Yards × 0.9144 = meters; . . . × 91.44 = centimeters.
Miles × 1.609 = kilometers; . . . × 1609 = meters.

Area:

Square inches × 0.0006452 = square meters; . . . × 6.452 = square centimeters.
Square feet × 0.09290 = square meters; . . . × 929.0 = square centimeters.
Square yards × 0.8361 = square meters; . . . × 8361 = square centimeters.
Acres × 4047 = square meters; . . . × 0.004047 = square kilometers; . . . × 0.4047 = hectares.
Square miles × 2,590,000 = square meters; . . . × 2.590 = square kilometers; . . . × 259.0 = hectares.

Volume:

Cubic inches × 0.01639 = liters; . . . × 16.39 = cubic centimeters.
Cubic feet × 28.32 = liters; . . . × 0.02832 = cubic meters.
Cubic yards × 764.6 = liters; . . . × 0.7646 = cubic meters.
Pints × 0.4732 = liters; . . . × 0.0004732 = cubic meters.
Quarts × 0.9463 = liters; . . . × 0.0009463 = cubic meters.
Gallon × 3.785 = liters; . . . × 0.003785 = cubic meters.
Acre-feet × 1233 = cubic meters.
Million gallons × 3,785,000 = liters; . . . × 3785 = cubic meters.

Weight or mass:

Grains × 0.06480 = grams; . . . × 0.00006480 = kilograms.
Ounces (avoirdupois) × 28.35 = grams; . . . × 0.02835 = kilograms.
Pounds (avoirdupois) × 453.6 = grams; . . . × 0.4536 = kilograms.
Tons (short) × 907.2 = kilograms; . . . × 0.9072 = metric tons.
Tons (long) × 1016 = kilograms; . . . × 1.016 = metric tons.

Specific combinations:

Feet per second × 1.097 = kilometer per hour; . . . × 0.3048 = meters per second; . . . × 0.5921 = knots.
Miles per hour × 1.609 = kilometers per hour; . . . × 0.4470 = meters per second; . . . × 0.8684 = knots.
Pounds per square inch × 70.3 = grams per square centimeter.
Pounds per square foot × 0.4885 = grams per square centimeters.
Tons (short) per square foot × 0.9765 = kilograms per square centimeter.
Tons (short) per acre × 0.2241 = kilograms per square meter; . . . × 2241 = kilograms per hectare.
Tons (short) per square mile × 0.0003502 = kilograms per square meter; . . . × 350.2 = kilograms per square kilometer.
Pounds per cubic foot × 0.01602 = grams per cubic centimeter; . . . × 16.02 = kilograms per cubic meter.
Cubic feet per second × 1.699 = cubic meters per minute; . . . × 0.02832 = cubic meters per second.
Cubic feet per second for 1 day × 1.983 = acre feet; . . . × 2446 = cubic meters.
Degrees Fahrenheit −32 × 0.556 = degrees Celsius.

FLUVIAL SEDIMENT CONCEPTS

By Harold P. Guy

Abstract

This report is the first of a series concerned with the measurement of and recording of information about fluvial sediment and with related environmental data needed to maintain and improve basic sediment knowledge. Concepts presented in this report involve (1) the physical characteristics of sediment which include aspects relative to weathering, soils, resistance to erosion, and particle size, (2) sediment erosion, transport, and deposition characteristics, which include aspects relative to fine sediment and overland flow, coarse sediment and streamflow, variations in stream sediment concentration, deposition, and denudation, (3) geomorphic considerations, which include aspects relative to the drainage basin, mass wasting, and channel properties, (4) economic aspects, and (5) data needs and program objectives to be attained through the use of several kinds of sediment records.

Introduction

It has long been the desire of hydrologists, hydraulic engineers, and others to develop a set of "universal" equations that would make it possible to predict the amount and characteristics of sediment erosion, transport, and deposition. Just as streamflow or groundwater predictive equations are still far from complete, it can be expected that there is only a very remote possibility for the development of a set of general equations to predict the many aspects of sedimentation.

The purpose of this chapter on "Fluvial Sediment Concepts" is to provide some knowledge of fluvial sedimentation and its implications in order that the reader can better understand why additional sediment data are needed and so that he can better decide where to make what kind of measurements. To this end, the subjects of weathering and soil formation, erosion resistance, and particle size are discussed with respect to the physical characteristics of sediment; fine sediment and overland flow, coarse sediment and streamflow, variations in concentration of sediment, and deposition are discussed with respect to erosion and transport; the drainage basin, mass wasting, and channel properties are discussed with respect to geomorphic aspects; some economic aspects are presented; and data needs and program objectives for several kinds of records are discussed.

Fluvial sedimentation includes the processes of erosion, transport, and deposition of soil or rock fragments. In conjunction with other forces, these natural phenomena have provided the major features of our landscape and channel systems as we see them today. Most sediment problems are related to one or more of three aspects: (1) Accelerated erosion because of poor land-use practices involving improper management in agriculture, in construction, and in the use of natural and manmade water courses, (2) stream erosion and deposition that affect specific kinds of land and water use, and (3) esthetic or physical damage by suspended sediment for many uses of water.

The conversation, development, and utilization of our land and water resources will always involve sedimentation problems to some degree. Many human activities, for example, increase or reduce the amount of runoff water, concentrate its flow, and (or) alter the natural resistance to flow and sediment movement. Such changes in the amount of natural flow and in the conveyance systems are the key to sediment problems. One might think that the solution to sediment problems would be to stop erosion. This is physically and economically impossible;

moreover, such activity would upset the present environment and cause many new problems, which in aggregate might be worse than the original sediment problems. In instances where some control of sediment may be desirable to alleviate a problem, the best solution may not be possible because the source of the problem may be at a location where controls cannot be applied as a result of legal and institutional constraints.

As noted by Gottschalk (1965, p. 264), it is evident that much new knowledge is still needed relative to the many aspects of erosion, transport, and deposition of sediment before predictions can be made regarding what will happen when a set of environmental conditions is altered. This chapter presents sediment concepts that should make it possible to obtain more useful measurements of the amount and nature of sediment involved in or interfering with desirable utilization of our land and water resources. Because of the extensive condensation of the literature used to present these concepts, it is expected that the reader may find it necessary to obtain further detail from the listed references, and others, in order to complete the comprehensive picture on fluvial sediment and to help cope with some of the problems with special measurements.

The author acknowledges with warm appreciation the encouragement and helpful suggestions and criticisms from many colleagues. Particular thanks are extended to S. K. Love and W. H. Durum, former and present chiefs of the Quality of Water Branch, for their encouragement, and to F. C. Ames and D. M. Culbertson for their technical assistance. Many helpful comments have also been received from C. R. Collier, R. F. Flint, R. F. Piest, L. A. Reed, and K. F. Williams.

Physical characteristics

The principal source of fragmental material that may become fluvial sediment is the disintegration of rocks of the earth's crust. Such disintegration is for the most part caused by several physical and chemical weathering processes. As a result, and perhaps as a part of the weathering processes, soils are formed that have widely varying characteristics depending on climate, organisms, topography, parent material, and time. The erodibility of such soils, or conversely their resistance to becoming fluvial sediment when exposed, depends not only on the physical size of the particles, but also on the nature of inorganic and organic materials that bind the particles together.

When eroded from the surface of the land or the channel bed or banks, the sediment or fragmental material may move rather continuously with the flow or be transported and deposited many times by the flow, the motion depending on the strength of the fluid forces in relation to the weight or resisting force of the particles. Once sediment particles are eroded, then the resistance to transport is directly related to the fall velocity or "fall diameter" of the particle. Concepts relating physical size to fall velocity must also include consideration of particle shape and specific gravity.

Weathering and soil formation

The four factors that affect the type and rate of rock weathering are rock structure, climate, topography, and vegetation (Thornbury, 1954, p. 37). Rock structure is characterized by many physical and chemical properties. Temperature and moisture are the important climatic factors that determine the kind and rate of weathering. Topography affects the exposure of rock to precipitation, temperature, and vegetation as well as to the forces of moving fluids. Decaying organic matter from vegetation produces carbon dioxide and humic acids that can attack rock.

According to Reiche (1950), the important physical processes that lead to rock fragmentation include: (1) expansion resulting from unloading, (2) crystal growth, (3) thermal expansion, (4) organic activity, and (5) colloid plucking. With respect to crystal growth, local formation of ice crystals by repeated freeze and thaw is a most effective weathering process in the middle and high latitudes during fall and spring. The pressure attained upon freezing of the interstitial water depends on how completely the water is confined. The ice-crystal weathering process should not be confused with frost heaving caused by the accumulation of ice masses within soils capable of rapid capillary movement of moisture.

It is generally recognized that chemical weathering is more important than physical weathering (Thornbury, 1954, p. 41). Chemical weathering includes hydration, hydrolysis, oxidation, reduction, carbonation, and solution. Chemical weathering often causes (1) an increase in bulk due to physical stresses within the rocks, (2) a change to smaller and more stable sizes of particles, and (3) the formation of lower density materials. Chemical weathering progresses toward the formation of those minerals that are in equilibrium at the surface of the earth. Relative mineral stability, as indicated by Goldich (1938), is given in the list below: from least to most stable is from top to bottom.

Olivine
Calcic plagioclase
Augite
Calci-alkalic plagioclase
Hornblende
Alkali-calcic plagioclase
Alkalic plagioclase
Biotite
Potash feldspar
Muscovite
Quartz

Thus, it is evident that quartz and muscovite should be the most common residual fragments of weathered rocks.

The following (Lyon and Buckman, 1943) summarizes in a rather simplified way the complex interrelationships of the weathering processes involved in the development of soil material from bedrock. The process is initiated by a physical weakening, often due to temperature changes, accompanied by chemical transformations involving hydrolysis and hydration of such minerals as feldspar, mica, and hornblende. The minerals thereby soften, lose their luster, and increase in volume. The colors in the decomposing mass are generally subdued, except for yellow or red caused by the formation of hematite or limonite. Cations released as a result of these changes, such as calcium, magnesium, sodium, and potassium undergo carbonation and are easily removed as water is drained away. Ultimately, all but the most resistant of the original minerals are removed leaving secondary hydrated silicates that often recrystallize into colloidal clay. A small amount of such clay results in a sandy, rather friable soil material, but when the clay is dominant, the mass is heavy and plastic.

Lyon and Buckman (1943) further emphasize that the rate of activity among the various weathering processes will be governed by climate. The soil material will more likely be coarse under arid conditions, where the physical forces may dominate, and higher colloidality and finer materials can be expected in the humid regions, where all processes are involved, especially the vigorous chemical changes. Also, the forces of weathering lose their intensity with depth below the surface; moreover, the transformations are likely to be different because of larger amounts of water and a decrease in porosity and aeration. Such differences with depth result in the formation of a definite soil profile from the decomposing mass of rock materials.

Some additional explanation of the basic soil-forming process is essential to a better understanding of the nature of sediments available for fluvial processes and of their resistance to erosion. Soil can be defined in a number of ways; the definition patterned after that of Bushnell (1944) is appropriate. Soil is a natural part of the earth's surface and is characterized by layers, roughly parallel to the surface, formed in time by physical, chemical, and biological processes operating on parent materials.

Soil classification once was highly dependent on geology and was concerned with whether or not the parent material was residual or transported; it is now more dependent on the chemical and physical characteristics of the successive layers that constitute the soil profile. A matured soil profile has an A horizon or layer immediately beneath the surface. This layer is eluvial or leached; that is, solutes and fine clays have been removed by descending soil water and organic materials may be accumulated. (See fig. 1.) The B horizon, commonly called subsoil, is an illuvial or "washed in" layer where solutes have been precipitated and the clays from the A horizon have been trapped. The C horizon is the parent material or the partially weathered rock products not seriously affected by the movement of soil water. It is therefore evident that the characteristics of a young soil would be close to those of the parent material whereas the characteristics of a mature soil would be more closely related to climate and vegetation.

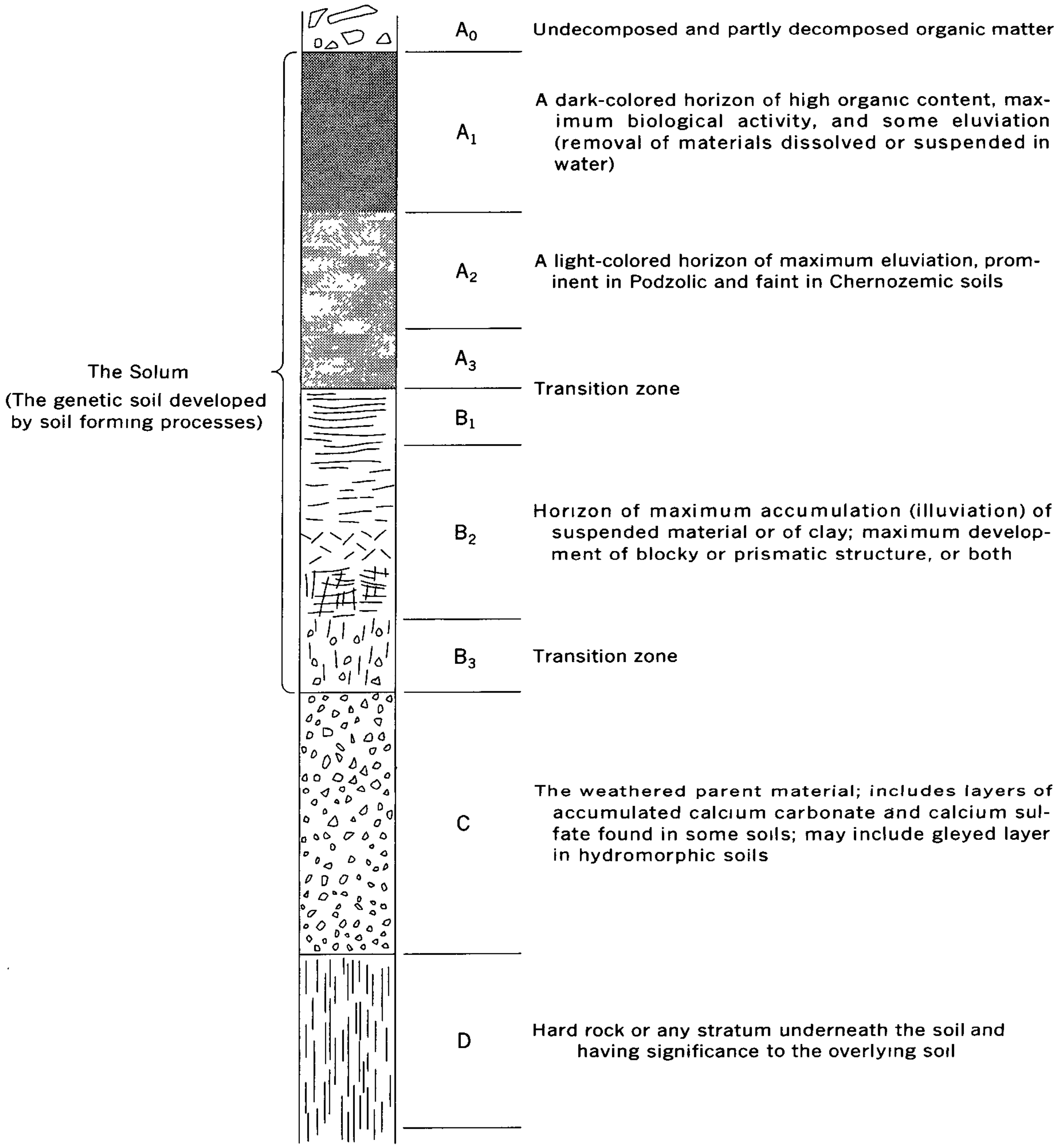


Figure 1.—A hypothetical soil profile of the principal horizons. Every profile has some, but not all, of the indicated features. Modified from Simonson (1957, p. 20).

The recognized soils of the world can be included in ten orders (U.S. Department of Agriculture—Soil Conservation Service, 1960) in a classification now extensively used by such agencies as the Soil Conservation Service. The present orders can best be introduced by relating them to the kinds of soil recognized in previous classifications as indicated in table 1. The previous classifications (Lyon and Buckman, 1943; U.S. Department of Agriculture, 1938) were based mainly on climatic and vegetative conditions as well as the degree of weathering

Table 1.—Present soil orders and approximate older equivalents

[Derivation of element: L., Latin; Gk., Greek; F., French]

Present order	Formative element in name	Derivation of element	Approximate equivalents
1. Entisols	ent	Nonsense syllable, recent	Azonal soils, and some Low-Humic Gley soils.
2. Vertisols	ert	L. verto, turn	Grumusols.
3. Inceptisols	ept	L. inceptum, beginning	Ando, Sol Brun Acide, some Brown Forest, Low-Humic Gley, and Humic Gley soils.
4. Aridisols	id	L. aridus, dry	Desert, Red Desert, Sierozem, Solonchak, some Brown and Reddish Brown soils, and associated Solonetz.
5. Mollisols	oll	L. mollis, soft	Chestnut, Chernozem, Brunizem (Prairie), Rendzina, some Brown, Brown Forest, and associated Solonetz and Humic Gley soils.
6. Spodosols	od	Gk. spodos, wood ash	Podzols, Brown Podzolic soils, and Ground-Water Podzols.
7. Alfisols	alf	Nonsense syllable, pedalfer.	Gray-Brown Podzolic, Gray-Wooded soils, Noncalcic Brown soils, Degraded Chernozem, and associated Planosols and some Half Bog soils.
8. Ultisols	ult	L. ultimus, last	Red-Yellow Podzolic soils, Reddish-Brown Lateritic soils of the United States, and associated Planosols and Half Bog soils.
9. Oxisols	ox	F. oxide, oxide	Laterite soils, Latosols.
10. Histosols	ist	Gk. histos, tissue	Bog soils.

and particle movement. The ten orders and a partial description of each follow:

1. **Entisols** at one extreme in age might consist of very recent alluvium, perhaps with gray or brown mottling in the epipedon—some mottles can develop in alluvium before the floodwaters that laid down the deposit have receded. At the other extreme in age, Entisols may include quartz sands in place for many thousands of years. Under certain conditions quartz sands may form Humaquods or Humods. In summary, Entisols are composed of deep regolith with no definite horizons except a plow layer. Their color ranges from the bluish gray of tidal marshes through blacks, grays, yellows, browns, and reds. In arid lands, they may contain small accumulations of carbonates, sulfates, or other more soluble salts, but not enough to constitute calcic, gypsic, or salic horizons.
2. **Vertisols** include the swelling clays normally developed in montmorillonitic parent materials derived from limestone or basic igneous rocks. Technically, Vertisols contain more than 35 percent expanding-lattice clay and more than 30 milliequivalents exchange capacity in all horizons more than 5 cm deep; at some seasons they contain cracks 1 to 25 cm wide that reach to the middle of the solum. The climate may range from subhumid to arid and from tropical to temperate. The natural vegetation of Vertisols is usually grass or herbaceous annuals, but sometimes scattered drought-tolerant woody plants may be present.
3. **Inceptisols** are found on young but not recent land surfaces and contain one or more rather quickly-formed horizons that do not represent significant illuviation or eluviation or extreme weathering. Included are many soils formerly called Brown Forest soils, Tundra, Lithosols, and Regosols, and a number of the strongly gleyed soils such as Humic Gley and Low-Humic Gley. Inceptisols may have notable textural differences between horizons only if parent materials are stratified. They are normally found in humid climates and range from the Arctic to the Tropics and to alpine areas under a native vegetation, most often a forest.
4. **Aridisols** include primarily the soils of places usually dry when not frozen and include those previously called Desert soils, Red Desert soils, Sierozems, Reddish

Brown soils, and Solonchak. Moist soils in dry places may be included that have no argillic or spodic horizon, but have a calcic, gypsic, or salic horizon.

5. **Mollisols** include most soils that have been called Chernozem, Prairie, Chestnut, and Reddish Prairie, the Humic Gley soils, and Planosols. The Mollisols must have a mollic epipedon but exclude those with a mollic epipedon dominated by allophane or a silt and sand fraction dominated by volcanic ash. Most have developed under a grass vegetation spaced closely enough to form a sod. A few have developed under hardwood forest where there are basic and calcareous parent materials and a large earthworm population.
6. **Spodosols** are formed on nonclayey siliceous parent materials, in humid regions from the boreal forests to the tropics, mostly under coniferous forest. They have been called Podzols, Brown Podzols, and Ground-Water Podzols. The main criterion is that a spodic horizon be present, though several other diagnostic horizons, such as histic, umbric, ochric, and argillic horizons and duripans and fragipans, may be found.
7. **Alfisols** are mineral soils, generally moist, with no mollic epipedon, or oxic or spodic horizon, and with an argillic or natric horizon. They include most soils that previously have been called Noncalcic Brown soils, Gray-Brown Podzolic soils, Gray-Wooded soils, some Planosols, and Half Bog soils. The requirement of a high base saturation in the argillic horizon suggests that there has been little movement of water through the soil or that the parent materials are young, unweathered, and basic. Therefore, in humid climates, the parent materials are generally no older than Pleistocene and contain carbonates.
8. **Ultisols** have an argillic but no oxic or natric horizon. They may have a mollic, umbric, ochric, or histic epipedon, or a fragipan and plinthite are often present. The Ultisols include most soils that have been called Red-Yellow Podzolic soils, Reddish-Brown Lateritic soils, and Rubrozems and some of the very acid Humic Gley and Ground-Water Laterite soils. They range from the temperate zones to the tropics, occur on land surfaces that are relatively old, and develop under forest, savannah, or even marsh or swamp flora. The exclusion of oxic horizons requires that some weatherable materials be present including small amounts of micas or feldspars in the silt and sand fraction and (or) allophane or 2:1 lattice clays.
9. **Oxisols** have oxic horizons and the epipedon may be umbric, histic, or perhaps mollic. Sometimes an argillic horizon may be present. They generally occur in the tropical and subtropical regions on old land surfaces and have been called Latosols and Ground-Water Laterites.
10. **Histosols** have previously been called Bog soils or organic soils and may include some Half Bog soils. Decomposition of organic materials results in a dark-colored surface layer of finely divided muck of varying thickness. They may have either a mollic epipedon of high base saturation, a pH of more than 5, and carbon-nitrogen ratios less than 17, or an umbric epipedon that has a pH less than 5 and carbon-nitrogen ratios of more than 17.

Erosion resistance

Aside from several kinds of mass wasting, the amount of a specific size or kind of sediment in a stream depends on the erosion of soils in the drainage basin and their transport to and within the stream channel system. Although wind, glaciers, and even groundwater may erode sediment, the most significant erosional agent is running water. Thornbury (1954, p. 47) states that erosion can result from the acquisition or plucking of loose fragments by the erosional agent, the wearing away of resistant surfaces by impact from materials in transit, and the mutual wear of particles in transit through contact with each other. It is further understood that, without transportation, erosion of a specific layer of soil cannot occur until the layer above has been removed.

As expected, the amount of erosion can be related to climate or to mean annual temperature and rainfall as indicated in figure 2. Erosion would be expected to be the least where the

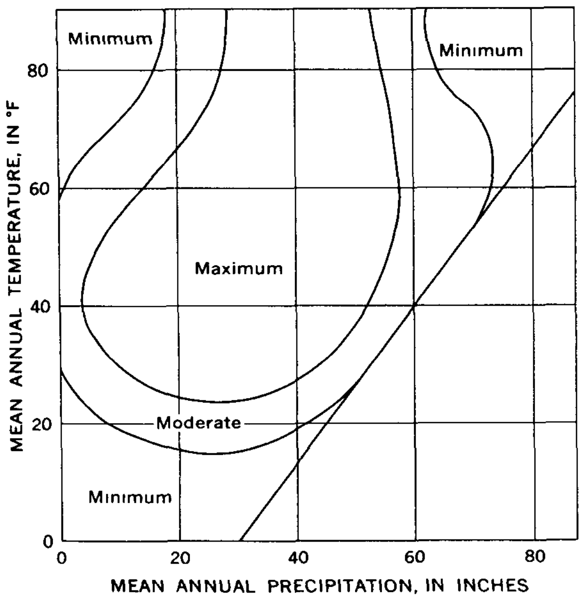


Figure 2.—Relative erosion as related to mean annual temperature and precipitation. Redrawn from Thornbury (1954, p. 60).

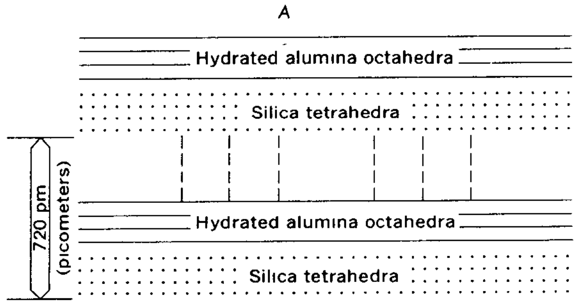
temperature is at or below freezing, where rainfall and temperature are adequate to produce dense vegetative cover, and where rainfall is insufficient at high temperatures to yield runoff because of evapotranspiration. Maximum erosion then occurs at combinations of precipitation and temperature that result in a combination of rapid weathering, maximum runoff, and relatively sparse vegetation. These factors imply also that for a given location and mean precipitation and temperature, a highly variable climate will cause more erosion than would a nonseasonal climate.

The active erosional agents are generally in balance with a set of resisting forces. Such resisting forces may include the gravitational and interlocking forces of the particles and the many kinds of organic and inorganic binding agents. Pure rock fragments, sands, and even silt-sized materials contain little or no binding agent and, therefore, must depend on the interlocking forces to resist erosion. Baver (1948) states that the silt and sand fractions may be considered as the skeleton of the soil in the absence of marked physical or chemical activity and that the clay and humus material are the active parts because of their chemical composition and high specific surface.

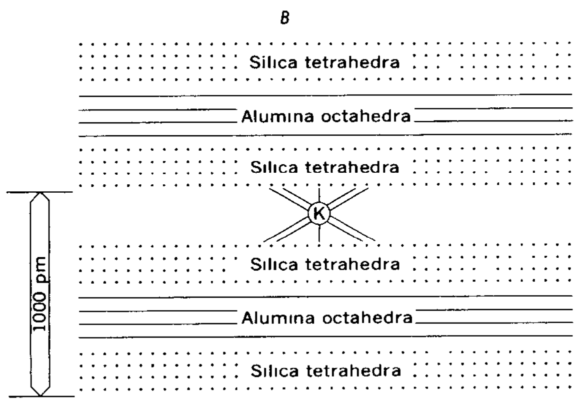
The tractive force required to move a particle against only the interlocking gravitational force can be computed from a hydraulic point of view. The binding forces, on the other hand, are of diverse character and operate by chemical reaction through association of a very large number of very small particles, generally less than 0.002 mm (millimeters). According to Russell (1957), clay minerals are secondary hydrated aluminosilicates in which isomorphous substitutions have occurred. Figure 3 shows the schematic arrangement of kaolinite, illite, and montmorillonite crystals.

Kaolinite, which is in most mature soils, consists of alternating silicon-oxygen and aluminum-oxygen layers (Al:Si::1:1) in double-layered sheets joined by hydrogen bonds. The space between the double-layered sheets is "fixed" and inaccessible for surface reactions. Illite and montmorillonite, on the other hand, have silicon-oxygen and aluminum-oxygen layers bonded together in a 2:1 ratio, thus making it possible for Al^{+3} to be substituted for Si^{+4} and Mg^{+2} or Fe^{+2} to be substituted for Al^{+3}. Such substitution may give the crystal a negative charge, in which case reaction with other charged particles and ions and with dipolar molecules such as water may occur. Thus illite and montmorillonite have considerable "exchange capacity." It is also noted that the charged clay surfaces can cause layers of water molecules at the surfaces to become oriented, and this gives the characteristic properties of plasticity, cohesion, and shrinkage to clays and soils that contain a large amount of the 2:1 lattice clay.

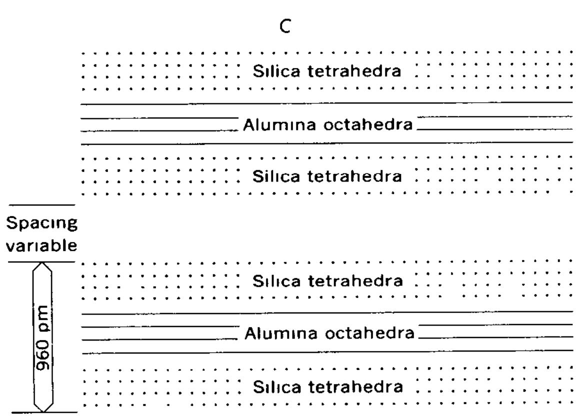
A soil aggregate consists of a grouping of a number of primary particles into a secondary unit. Flocculation occurs when primary particles attract each other upon collision in a water suspension with a low electrokinetic potential. Most such floccules are unstable and break up in other suspensions that lack the required flocculating agent. Baver (1948) states that stable-aggregate formation in soils requires that the primary particles be so firmly held together that they do not readily disperse. Most of the cementing agents for stable-aggregate formation are the irreversible or slowly reversible inorganic colloids, such as the oxides of iron



Kaolinite crystals are composed of pairs of silica and alumina sheets held together by hydrogen bonds. The space between the crystal units is fixed and is largely inaccessible for surface reactions



The crystal unit of illite consists of a silica sheet on each side of an alumina sheet. Adjacent crystal units are held together by potassium bridges The space between the units is partly accessible for surface reactions



The crystal unit of montmorillonite consists of a silica sheet on each side of an alumina sheet. The interlattice spacing in the montmorillonite clays varies with the amount of water present The entire surface of the crystal unit is accessible for surface reactions

Figure 3.—Schematic arrangement of clay minerals: (A) kaolinite, (B) illite, and (C) montmorillonite. Redrawn from Russell (1957, p. 33–34).

and alumina, and the organic colloids. Organic colloids are the intermediate products in the decomposition of plant residues; they are adsorbed on the surface of soil particles through hydrogen bonding. The strength of the colloid bond is increased if irreversible dehydration and shrinkage occurs.

Aggregate analysis of a large number of different soils has shown that there is a strong correlation between climate and aggregation. (See fig. 4.) The percentage of aggregates is at a maximum in the semiarid and semihumid regions. Aggregation is low in Desert soils because of small clay content, which in turn is caused by slow and incomplete chemical weathering. Aggregation is also low in the Podzols because the climatic forces have been sufficiently great to cause leaching of the colloids as they are formed.

The previous paragraphs illustrate why erosion is more complicated than merely lifting and moving fragmental sediment particles from a pile of such particles. The problem is also not one of forces that are constant and simply bind

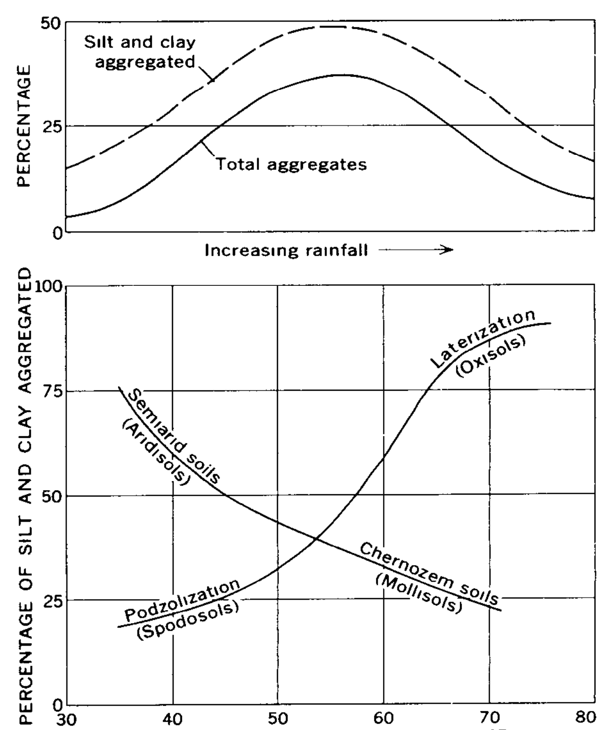


Figure 4.—Relationship of soil aggregation to climatic factors. From Baver (1948, p. 150).

particles together but is one of forces that change because of reactions with water and other ions. Soil erodibility is related to the many physical characteristics of soil that affect its resistance to erosion, but potential soil erosion includes potential erosivity and vegetative cover protectivity. Potential erosivity as used by Cook (1936) included the impact energy of raindrops, the infiltration and storage capacities of soil, and the steepness and length of slope. Therefore a nonerodible soil may not result in less erosion than an erodible soil on the same slope. For example, a dense nonerodible clay soil may produce more erosion than an erodible loose sandy soil on the same slope—the higher erodibility of a sandy soil may be counteracted by its greater infiltration capacity.

Particle size

Except for the finer sizes that form aggregates, single-particle motion characterizes the processes of erosion, transportation, and deposition of sediment. Clay-sized particles may form rather flat aggregates or floccules of particles and thus, as fluvial sediment, behave similarly to larger discrete particles. Coarser particles tend to be less flat, but still are far from spherical.

Because of the irregular shape and the variation in specific gravity, physical size is not a good index of the fluvial character of sediment particles. The dynamic properties of a particle up to about 2 mm can best be described by its fall velocity (U.S. Inter-Agency Report, 1957), which is a function of its volume, shape, and specific gravity and the viscosity and specific gravity of the fluid (water).

If particle-size data of sediment particles are to be comparable, then a **standard fall velocity** is required. This is defined as the average rate of fall that a particle would attain if falling alone in quiescent distilled water of infinite extent at a temperature of 24°C. A particle is assumed to reach its most stable orientation and reach an average terminal rate of fall in a short time after release. According to Stringham, Simons, and Guy (1969), some particles, at least of the larger sizes, have been found to be unstable. The fall of extremely fine particles in the range of Stokes' law is likely to be stable, although some variation in the net downward movement may be expected because the fluid cannot be made completely quiescent.

Fall velocity can logically be converted to a diameter-of-particle concept or hydraulic size, though it may be only an approximation of physical size. The **fall diameter** of a particle is defined as the diameter of a sphere with specific gravity of 2.65 that would have the same standard fall velocity as the particle. Thus, a given particle has only one fall diameter as determined by its resistance to fall in the fluid against the force of the earth's gravity.

A **standard sedimentation diameter** concept further requires the use of the standard fall velocity and the specific gravity of the particle. So defined, the standard sedimentation diameter depends only on the volume and shape of the particle. Also, its relation to nominal diameter depends on the effect of particle shape and roughness on the settling velocity of the particle in water at 24° C. Because there is only one standard sedimentation diameter for a particle, it is useful for comparing the effect of shape on the relations between nominal diameters, or even sieve diameters, and diameters which depend on fall velocity.

The **nominal diameter** of a particle is the diameter of a sphere that has the same volume as the particle (Lane, 1947). Nominal diameter generally implies an equivalent physical diameter; however, the concept can be associated with a sedimentation diameter because the sedimentation diameter is based on a spherical equivalent of the particle. This is especially true for the clay and silt particles that are too small (<0.062 mm) for easy physical size measurement. The sands from 0.062 to 2.0 mm may be measured either hydraulically or physically. The VA (visual-accumulation) tube is commonly used for the hydraulic measurement, and sieves, for the physical measurement. For these sands, it should be remembered that the nominal diameter is usually larger than the sieve diameter, the relative difference being greater at the smaller sizes. Particles of 4.0 mm and larger are usually measured physically by means of the sieves or by direct measurement for gravel, and by direct measurement only for sizes larger then gravel (64 mm). Direct physical measurement may be accomplished in one of two ways.

372-206 O—70——3

First, the longest, the intermediate, and the shortest mutually perpendicular axes can be measured directly, the average of which would represent the "diameter" of the particle; or second, the particle can be immersed in a liquid, and the volume of displaced liquid is then converted into an equivalent nominal diameter.

The **shape factor,** needed in order to estimate the hydraulic size from measurements of physical size, can be computed by one of several formulas based on the measurements of axes a, b, and c (longest, intermediate, and shortest). The ratio $c/\sqrt{ab}$ is most commonly used (Corey, 1949). Alger and Simons (1968) proposed that this ratio be modified by the ratio of the diameter of a sphere whose surface area is equal to that of the particle to the nominal diameter of the particle, d_a/d_n. As expected, this modification is not very practical because of the difficulty of obtaining the surface area of such irregular particles.

With respect to particle roundness, Williams (1966) found that the fall velocities of sharp-edged cylinders and disks were 8 to 28 percent less than the fall of their well-rounded counterparts where all other particle properties were held constant. Surface texture or roughness, on the other hand, caused only a minor reduction in the fall velocities of such disks and spheres.

Further discussion of these particle-size concepts and methods of particle-size measurement can be found in chapter C1 book 5 of this report series, entitled "Laboratory Theory and Methods for Sediment Analysis" (Guy, 1968).

Erosion, transport, and deposition

The amount of sediment moving in a stream at a given site and at a given time is a function of a complicated set of active and passive forces acting on the land surface of the drainage basin and throughout the channel system upstream from the site. These forces involve the erosion and transportation capacity of the seemingly inconsequential and largely unnoticed raindrop splash and the overland flow as it makes its way to stream channels by way of sheet and rill flow. The most noticeable and recognizable forces involve the transporting and bank-eroding power of the channel flow at high rates derived from the accumulated overland flow or from large quantities of groundwater flow. Table 2, partly derived from Johnson (1961), illustrates the general relationship of the many factors affecting the erosion and transport of sediment. The relationships of environmental factors to fluvial sediment are poorly understood because, for the most part, only small and generally unrelated segments of the problem have been studied. Fluvial sediment is also poorly understood because of the interrelationships among the many diverse environmental factors in the many climatic regions and geographic areas.

Fine sediment and overland runoff

Overland runoff, the surface flow resulting from precipitation excess, is the most dynamic agent causing erosion and the consequent transport of sediment, especially the finer sizes. Rainfall intensity, infiltration capacity, and water storage at the land surface are important controlling sedimentologic factors, and they may vary greatly with time and location over a drainage system. The precipitation, for example, may vary from a light drizzle in the winter months to a heavy downpour during the warm summer months in the temperate zone. The infiltration ranges from zero for impermeable surfaces to several inches per hour for a very sandy soil or through a forest floor with good duff and a permeable subsoil. Surface storage may range from one or two hundredths of an inch in an urban area to more than an inch for a contour-furrowed agricultural crop.

The mechanics of splash, sheet, and rill erosion

Of the several active and passive environmental forces (table 2) that affect erosion and transport of sediment, rainfall is considered to be the most dynamic and hence at times by far the most important. At the beginning of a rainstorm on a surface of erodible sediment, the impact of raindrops will cause an aerial suspension of both dry and wetted sediment particles. The proportion of wet splashed particles will increase as the surface becomes wet to the maximum depth of the impact crater. Sediment particles in aerial suspension have a net transport in the downslope direction by gravity and (or) the leeward direction by wind.

Table 2.—Factors affecting erosion and transport of sediment from land surface

[Modified from Johnson (1961)]

Major factors	Elements	Influence of elements on soil erosion
Agents and characteristics causing active forces		
Climate	Rainfall-runoff (intensity and duration).	Raindrop splash erosion: Breaks down aggregates, dislodges and disperses soil, and thereby seals the surface and increases precipitation excess. Imparts turbulence to sheet flow causing movement of larger particles.
		Flow erosion: Physical force due to pressure difference and impact of water dislodges, disperses, and transports. Intensity and duration affect rate of runoff after infiltration capacity is reached.
	Temperature	Alternate freezing and thawing: Expands soil, increases moisture content, and decreases cohesion. Thus dislodgment, dispersion, and transport are facilitated.
	Wind	Pressure difference and impact: Dislodges by force due to pressure difference and (or) impact.
Gravity		Elements of mass wasting: See page 35.
Agents and characteristics causing passive forces		
Soil character	Properties of the soil mass.	Granulation: Affects force required for dislodgment and transport.
		Stratification: Stratum of lowest porosity and permeability controls infiltration rate through overlying layers.
		Porosity: Determines waterholding capacity. Affects infiltration and runoff rates.
		Permeability: Determines percolation rate. Affects infiltration and runoff rates.
		Volume change and dispersion properties: Soil swelling loosens and disperses soil and thereby reduces cohesion and facilitates dislodgment and transport.
		Moisture content: Moisture reduces cohesion and lengthens erosion period by increasing the period of precipitation excess.
		Frost susceptibility: Determines intensity of ice formation and affects porosity, moisture content, and reduction in strength.
	Properties of the soil constituents.	Grain size, shape, and specific gravity: Determines force needed for dislodgment and transport, against force of gravity.
Topography	Slope	Orientation: Determines effectiveness of climatic forces.
		Degree of slope: Affects energy of flow as determined by gravity.
		Length of slope: Affects quantity or depth of flow. Depth and velocity affect turbulence. Both velocity and turbulence markedly affect erosion and transport.
Soil cover		Vegetative cover: All vegetative cover, whether alive or dead, protects the land surface in proportion to interception of raindrops by canopy and retardation of flow erosion through decreasing velocity of runoff, increasing soil porosity, and for live plants, increasing soil moisture-holding capacity through the process of transpiration.
		Nonvegetative cover: Open surfaces result in a minimum of surface protection and therefore maximum splash erosion, reduced infiltration, increased runoff, and maximum erosion. A paved surface affords maximum surface protection with zero erosion and highly efficient runoff and transport characteristics.

Rainfall impact tends to destroy soil aggregates and to consolidate the surface. The movement of particles and consolidation cause a sealing of the soil surface and a reduction in infiltration rate. The reduced infiltration increases the amount of precipitation excess and thus, on the land surface, locally creates a sheet of flowing water with erosive energy and transport capacity of its own. Such a sheet of flow is not likely to be extensive or of uniform thickness because of variations in infiltration rate and in the planeness of the surface. The impact of raindrops on the thin sheet flow causes a turbulent flow where one would ordinarily expect laminar flow. As stated by Stallings (1957, p. 64–65),

> Under certain conditions, raindrop impact can at times move stones as large as 10 mm in diameter when they are partially or wholly submerged in water. . . . Surface flow assists the downhill motion even though, if acting alone, it would not move them.

Excluding the effects of raindrop splash, erosion and transport of sediment are negligible under the conditions of laminar flow, but as the water from such laminar flow collects in rivulets and larger channels, the resulting energy of flow with increased scale and intensity of turbulence can be sufficient to carry heavy loads of sediment, especially fine particles. The important passive forces, therefore, tend to alter the the depth and velocity patterns of overland or surface flow. For example, the flow will be spread thinly and uniformly where the resistance to flow and cohesiveness of the soil prevent rilling of a relatively plane slope. The flow may be concentrated in many rivulets in areas where resistance is not uniform and where erosion can easily form small channels.

The difference between sheet-like or shallow flow and rill and channel flow in eroding and transporting sediment is considerable. The shallow flow moves rather slowly and, except when impacted by large raindrops, has a small amount of tractive force and a large amount of resistance (relative roughness) from the land surface. The rill and channel flow, on the other hand, is confined to a small area of resistance and has relatively great depths and hence large tractive force or gravity potential. The energy of such concentrated flow can, therefore, be sufficient to move sand, gravel, or even boulders. The "original" shallow flow erodes and transports mostly fine-grained sediment, the silts and clays, whereas the rill and other types of concentrated channel flow will carry not only the fine-grained load derived from the sheet flow but also the fine and coarse sediments that may be eroded from the bed and banks of the channels.

Some of the mechanics of splash and sheet erosion are exemplified in the formation and upslope movement of steps on steep loess-mantled slopes (Brice, 1958). These consist of "catsteps" or "terracettes" having rather bare scarps and sod-covered treads. Brice presents evidence that the steps originate as low sod scarps at the upslope edge of bare patches in the sod cover and that these scarps increase in height by upslope retreat caused by erosion of the soil from the downslope edge of the sod patch.

Sayre, Guy, and Chamberlain (1963) listed five environmental factors affecting the supply of sediment moved into and through a stream channel and, most applicable, the fine material contributed from the drainage area. They are:

1. The nature, amount, and intensity of precipitation.
2. The orientation, degree, and length of slopes.
3. The geology and soil types.
4. The land use.
5. The condition and density of the channel system.

These factors can operate either independently or in conjunction to deter or to advance the rate of erosion and transport. Precipitation, for example, if occurring at a low intensity and at ideal intervals, may advance the growth of vegetation and thereby increase the deterring influences. On the other hand, if the precipitation is intense and follows a drought or occurs on an area without vegetative cover, it is likely to cause a large amount of erosion. Because of the large variance and interrelation associated with the preceding list of factors, it is difficult to attain desirable spatial and temporal definition of the sediment erosion and transport characteristics in most drainage areas.

Rainfall characteristics

Wischmeier and Smith (1958), in a correlation of rainfall characteristics with erosion and soil-loss data, showed that an index consisting of the product of rainfall energy and the maximum 30-minute intensity of the storm is the most important measurable precipitation variable to explain storm-to-storm variation of soil loss from field plots. This concept is based on the fact that large, fast-falling raindrops with a large amount of kinetic energy will cause much splash erosion, thereby sealing the surface and increasing the amount of surface runoff. The maximum 30-minute intensity is also proportional to both the total quantity of rainfall and the average intensity for a storm. Values of Wischmeier's erosion index for the area of the United States east of 105° W are given in figure 5.

Wischmeier's erosion index R is defined as 0.01 of the summation of the product of the kinetic energy of rainfall, in foot-tons per acre, and the maximum 30-minute rainfall intensity,

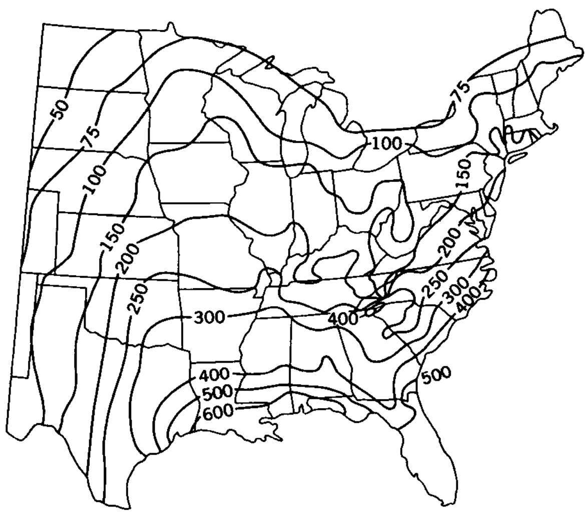


Figure 5.—Mean annual values of Wischmeier's erosion index for the area of the United States east of 105° W.

in inches per hour, for all significant storms on an average annual basis. This index has been found to be the most important measurable precipitation variable in the correlations with the storm-to-storm variation of soil loss from field plots.

Predicting sheet erosion

Data from field plot studies make it possible to develop general relationships for the prediction of erosion rates under a variety of land uses and environmental conditions. The following from Piest (1970) describes a commonly used equation:

The prediction model, known as the Universal Soil Loss Equation, was developed by Wischmeier, Smith and Uhland (1958) .... It has the general form

$$E=RKLSCP,$$

where E is the average annual soil loss, in tons/acre, from a specific field.

R is a rainfall factor expressing the erosion potential of average annual rainfall in the locality [fig. 5]. It is also called index of erosivity, erosion index, etc. The evolution of this parameter is traced by Wischmeier and Smith (1958).

K is the soil erodibility factor and represents the average soil loss, in tons/acre per unit of erosion index, R, from a particular soil in cultivated continuous fallow, with a standard plot length and percent slope arbitrarily selected as 73 feet and 9 percent, respectively. Pertinent values of the erodibility factor for a series of reference soils are obtained by direct measurement of eroded materials. Values of K for the soils studied vary from 0.02 to 0.70 tons/acre per unit of rainfall factor R.

S and L are topographic factors for adjusting the estimate of soil loss for a specific land gradient and length of slope [fig. 6]. The land gradient is measured in percent.

Slope length is defined as the average distance, in feet, from the point of origin of overland flow to whichever of the following limiting conditions occurs first: (1) the point where slope decreases to the extent that deposition begins or (2) the point where runoff enters well-defined channels.

C is the cropping management factor and represents the ratio of the soil quantities eroded from land that is cropped under specific conditions to that which is eroded from clean-tilled fallow under identical slope and rainfall conditions.

P is the supporting conservation practice factor (stripcropping, contouring, etc.). For straight-row farming, $P=1.0$.

A typical use for a sheet-erosion equation, as taken from a handbook based on Wischmeier and Smith (1965), might be to calculate the expected average annual soil loss from a given cropping sequence on a particular field. Consider a field in Fountain County, Ind., on Russell silt loam, having an 8-percent slope, a slope length of about 200 feet, and a 4-year crop rotation of wheat, meadow, and two seasons of corn. Assume that all tillage operations are on contour and that prior crop residues are plowed down in the spring be-

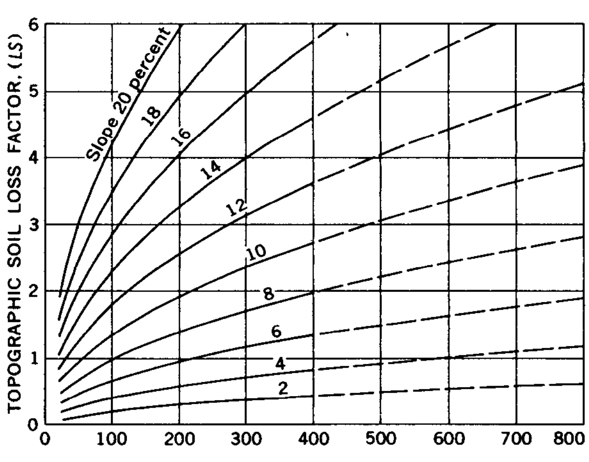


Figure 6.—Relationship of topographic soil-loss factor, LS, to slope length and gradient. The curves indicate that, for a given gradient, soil loss varies with the square root of the slope length.

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 soli-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 particles 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.

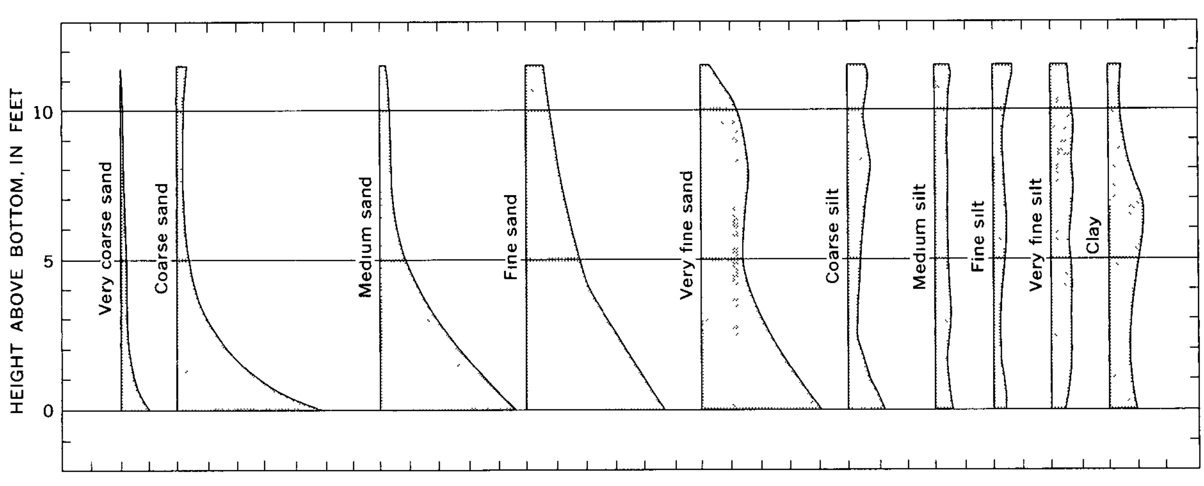


Figure 7.—Discharge-weighted concentration of suspended sediment for different particle-size groups at a sampling vertical in the Missouri River at Kansas City, Mo.

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-

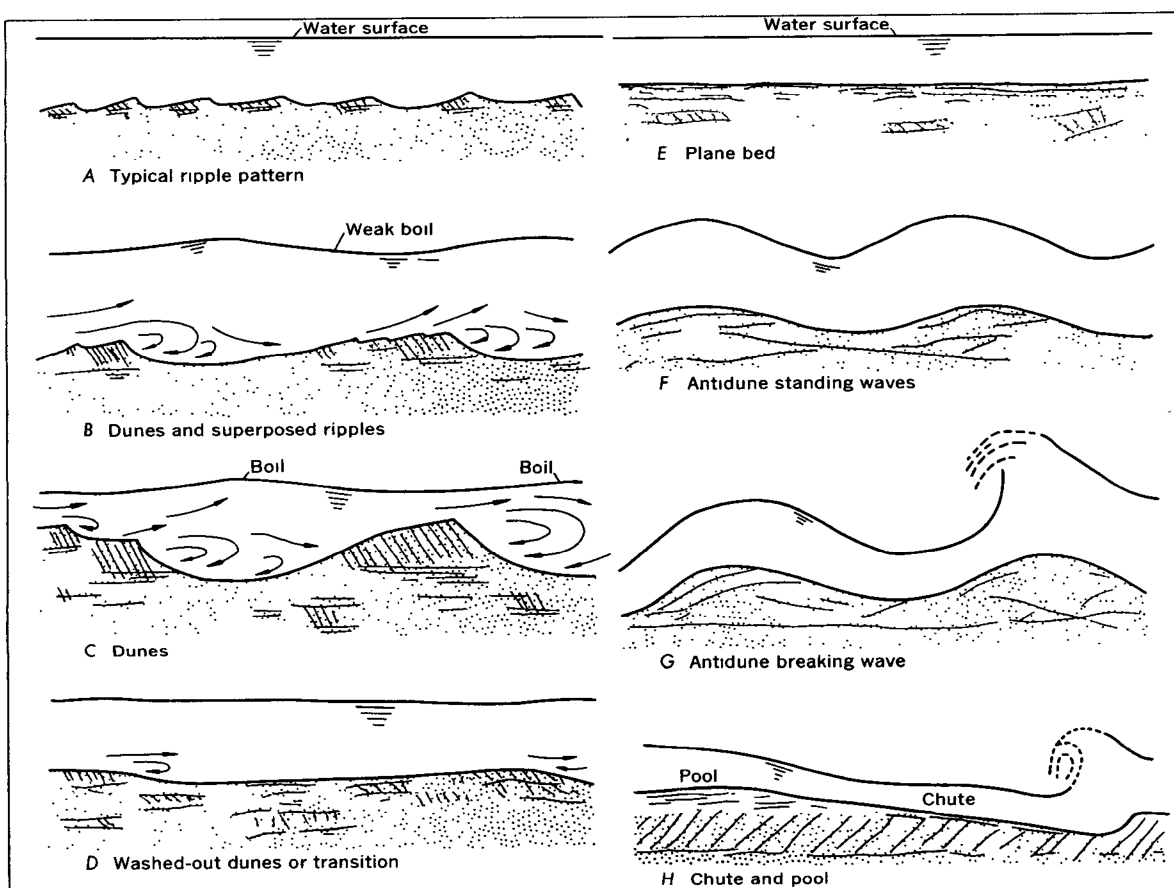


Figure 8.—Schematic diagrams of eight types of roughness found in sand-bed channels. Types A through C are representative of the lower flow regime where the Froude number is usually <0.4, E through H are representative of the upper flow regime where the Froude number is usually >0.7, and D represents the transition regime. Modified from Simons and Richardson (1966, p. J5).

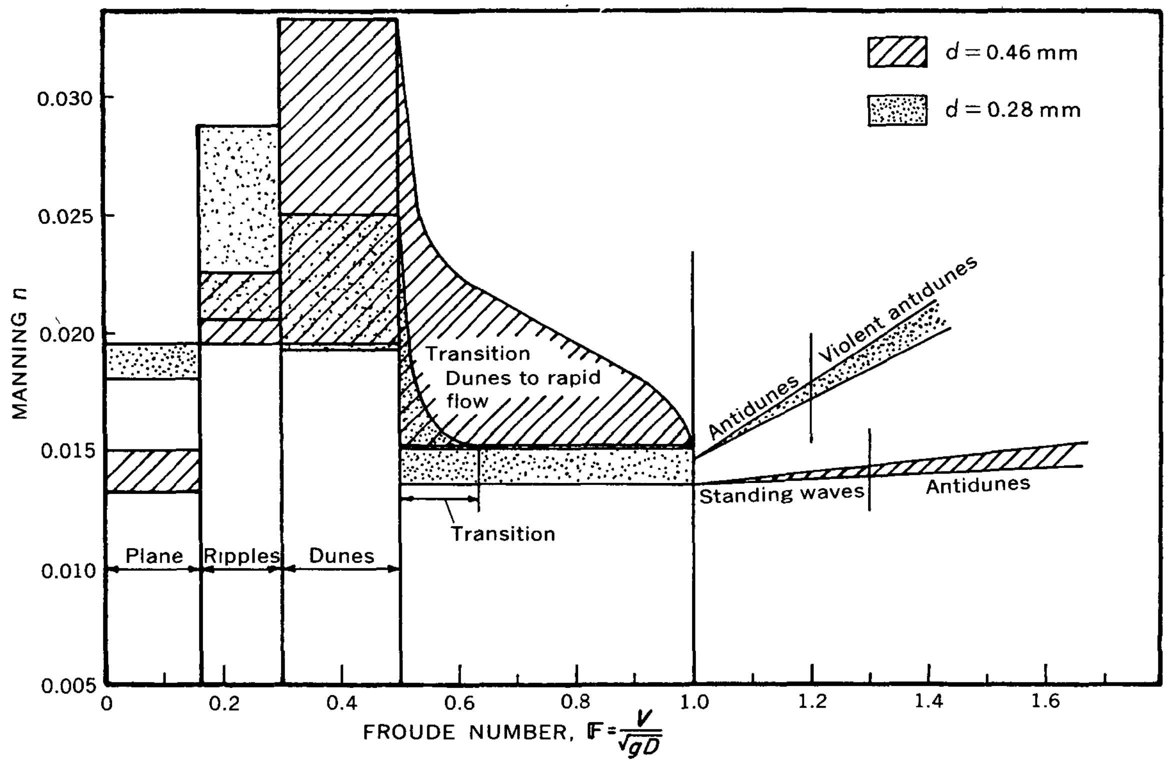


Figure 9.—Effect of size of bed material and Froude number on form of bed roughness and Manning n for a range of flow conditions with sands of 0.28- and 0.46-mm median diameter in an 8-foot-wide flume. Modified from Simons and Richardson (1962, p. E7).

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

372-206 O—70——4

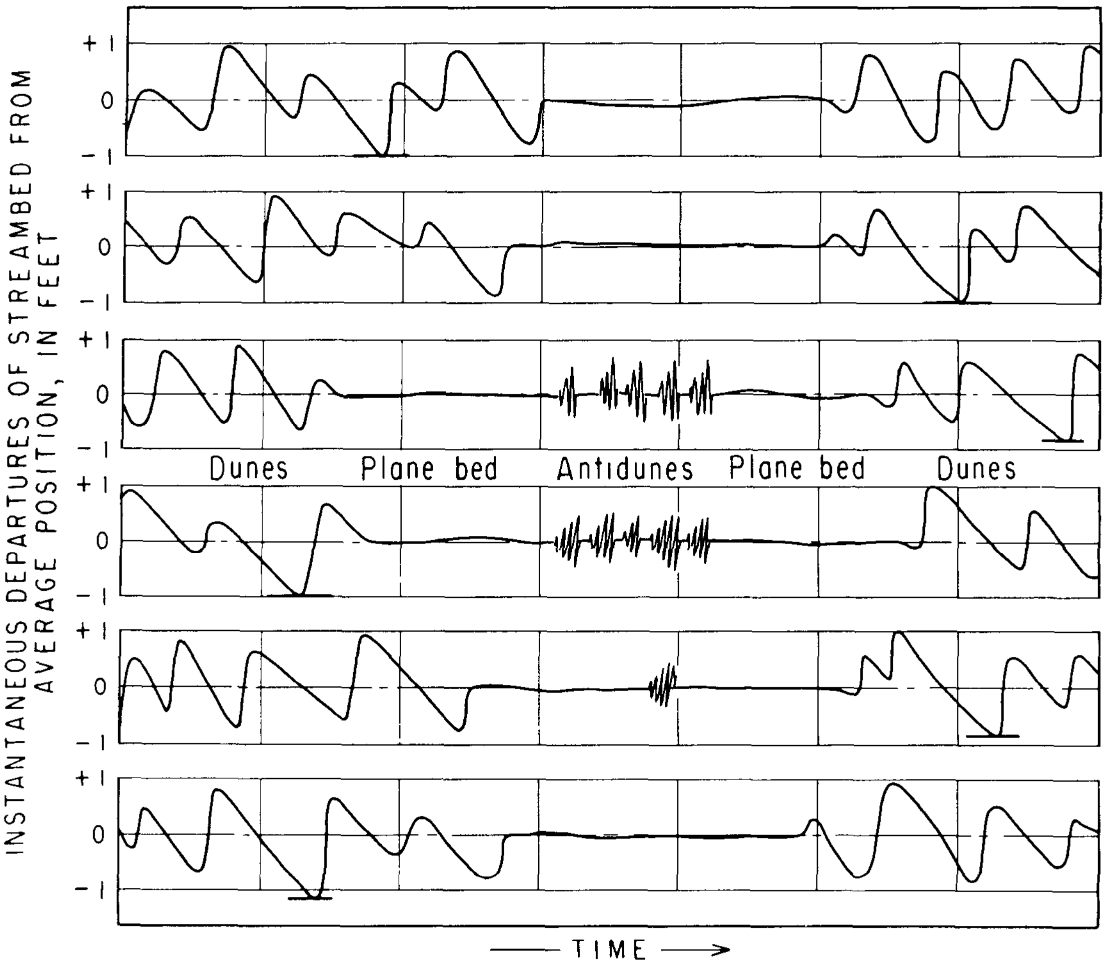


Figure 10.—Schematic diagram of streambed elevation with time at six points in a stream cross section. Time of changes not to scale. From Colby (1964b, p. 4).

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 sediment 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 proportional 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

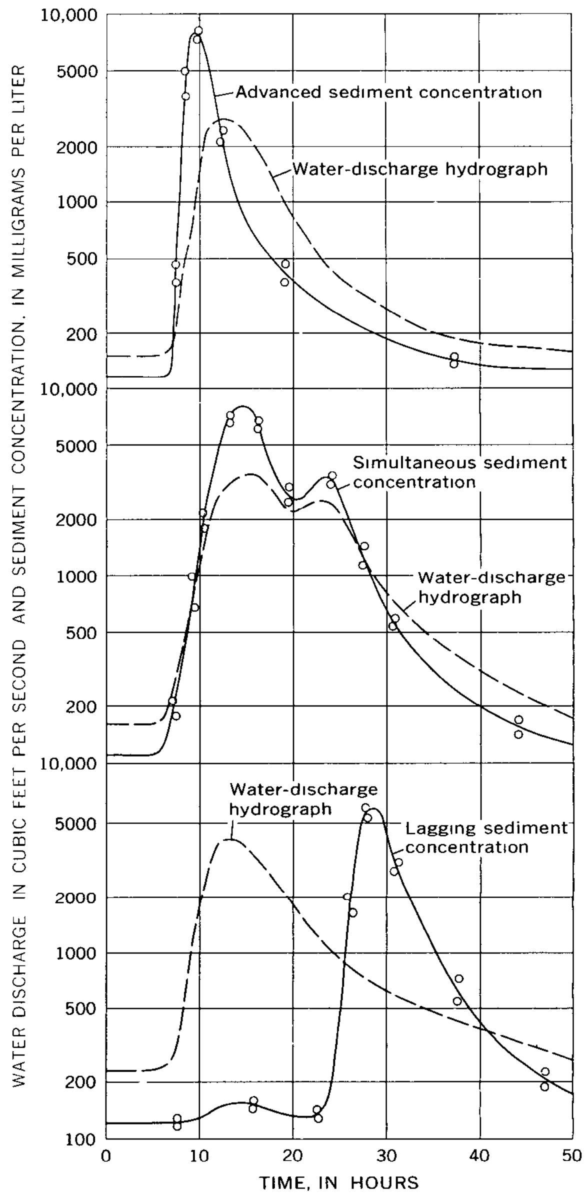


Figure 11.—Advanced, simultaneous, and lagging sediment-concentration graphs as related to the temporal distribution of their respective water-discharge hydrographs. Terms leading, inphase, and delayed are sometimes used.

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

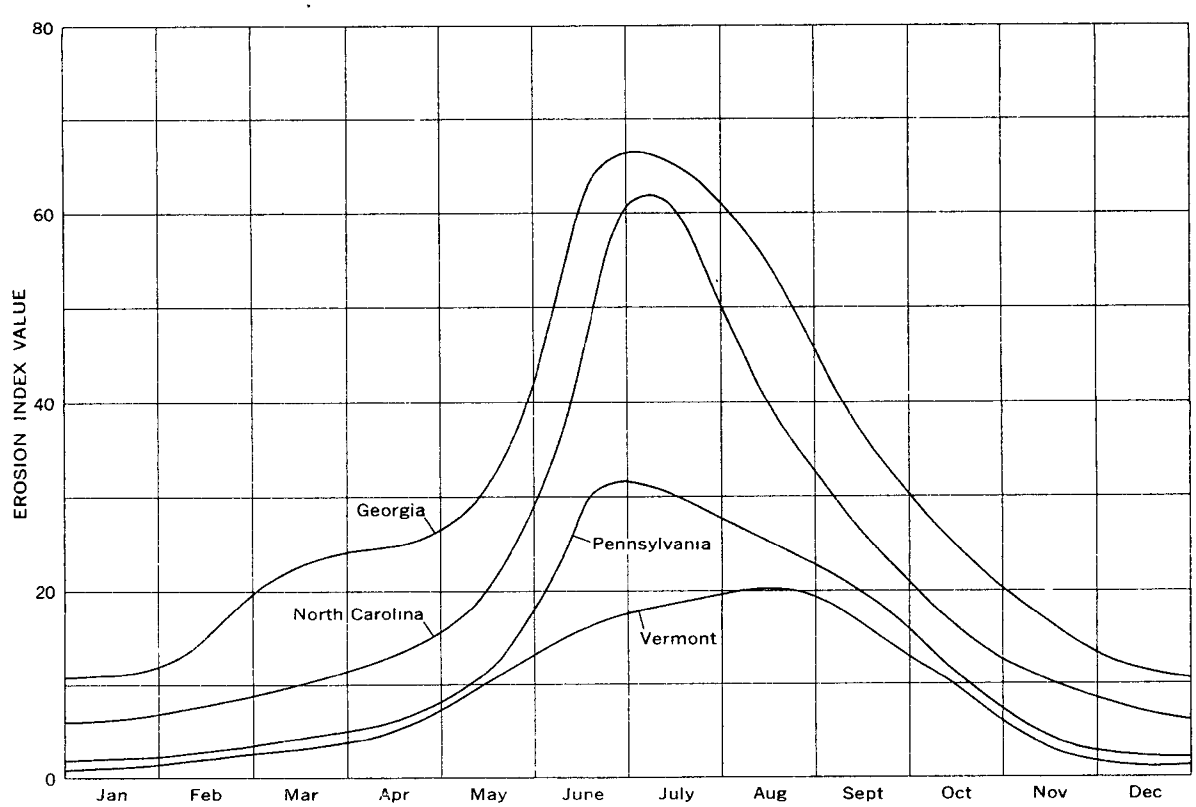


Figure 12.—Seasonal distribution of Wischmeier's erosion-index values at four locations in the Atlantic coast area. From Guy (1964, p. 10).

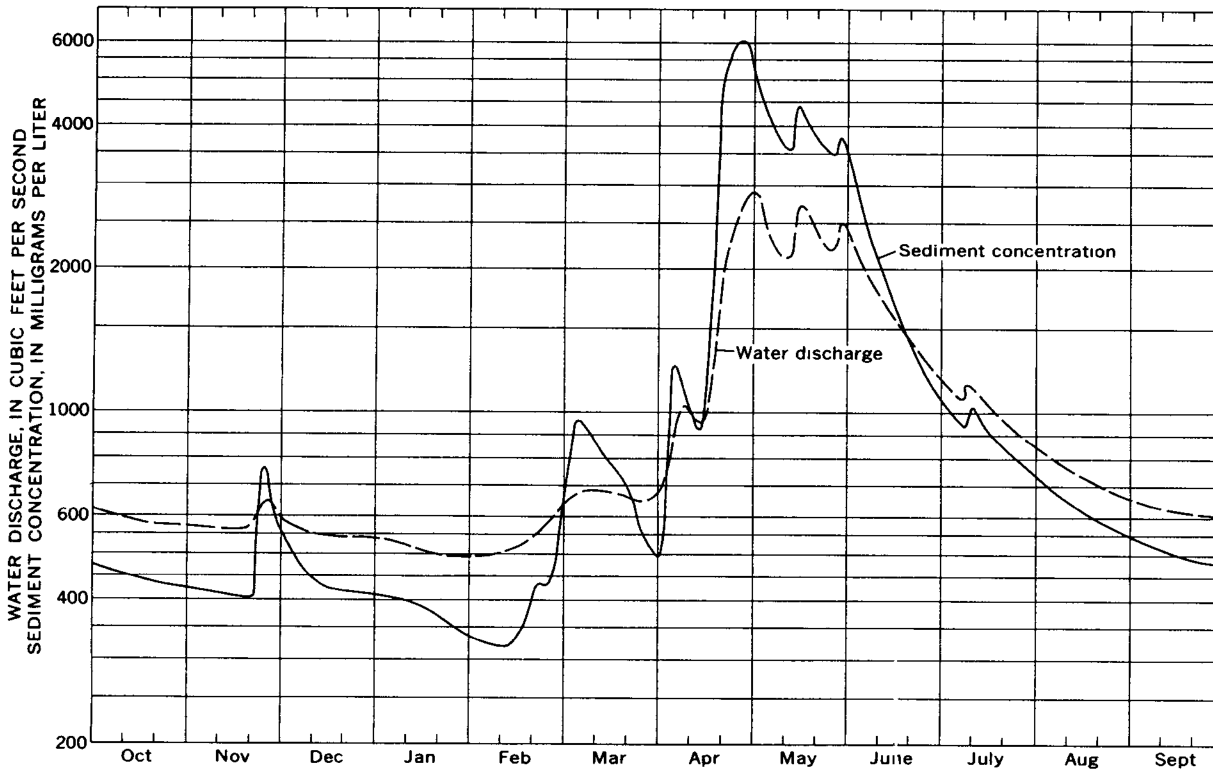


Figure 13.—Temporal relationship of sediment concentration to water discharge for an assumed "snowmelt" stream draining mountainous terrain.

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-discharge 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)

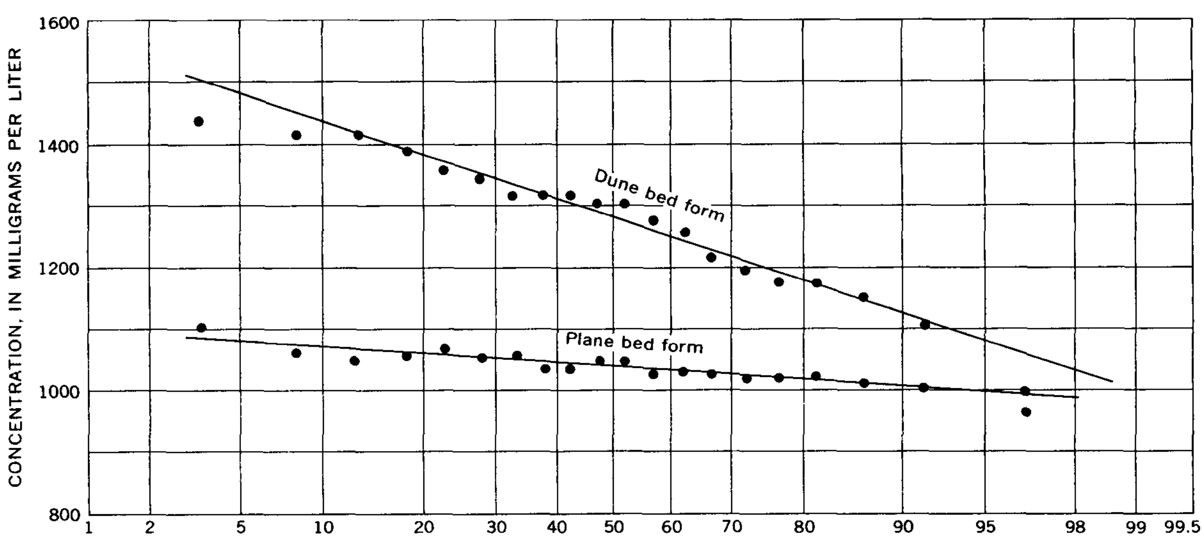


Figure 14.—Frequency distributions of consecutive sampled concentrations at single verticals of the Middle Loup River at Dunning, Nebr.

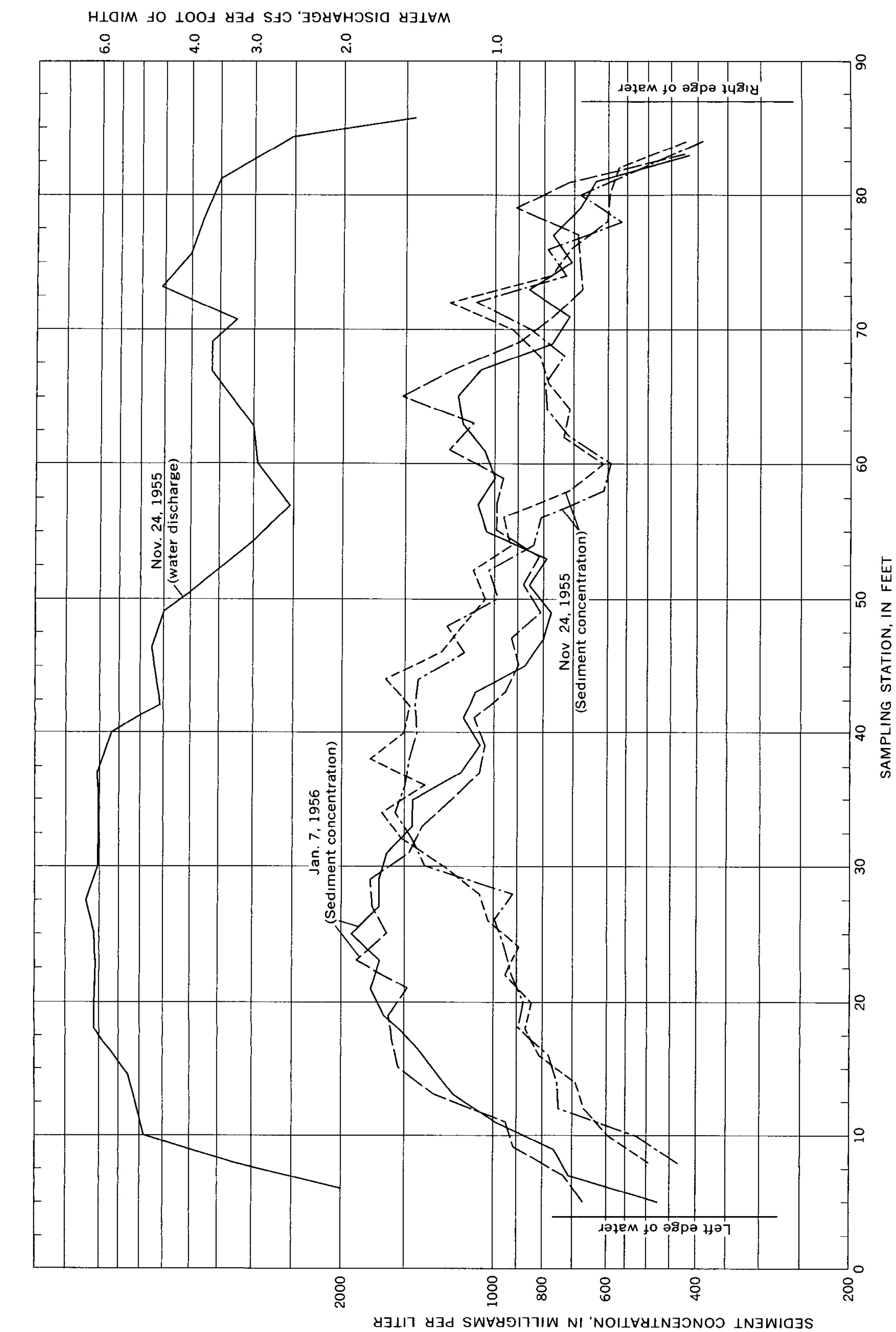


Figure 15.—Lateral variation of sampled sediment concentration for two different days on the sand-bed stream, Middle Loup River at Dunning, Nebr. The two measurements on each day were taken only a few minutes apart. Modified from Hubbell (1960).

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 streamflow, 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:

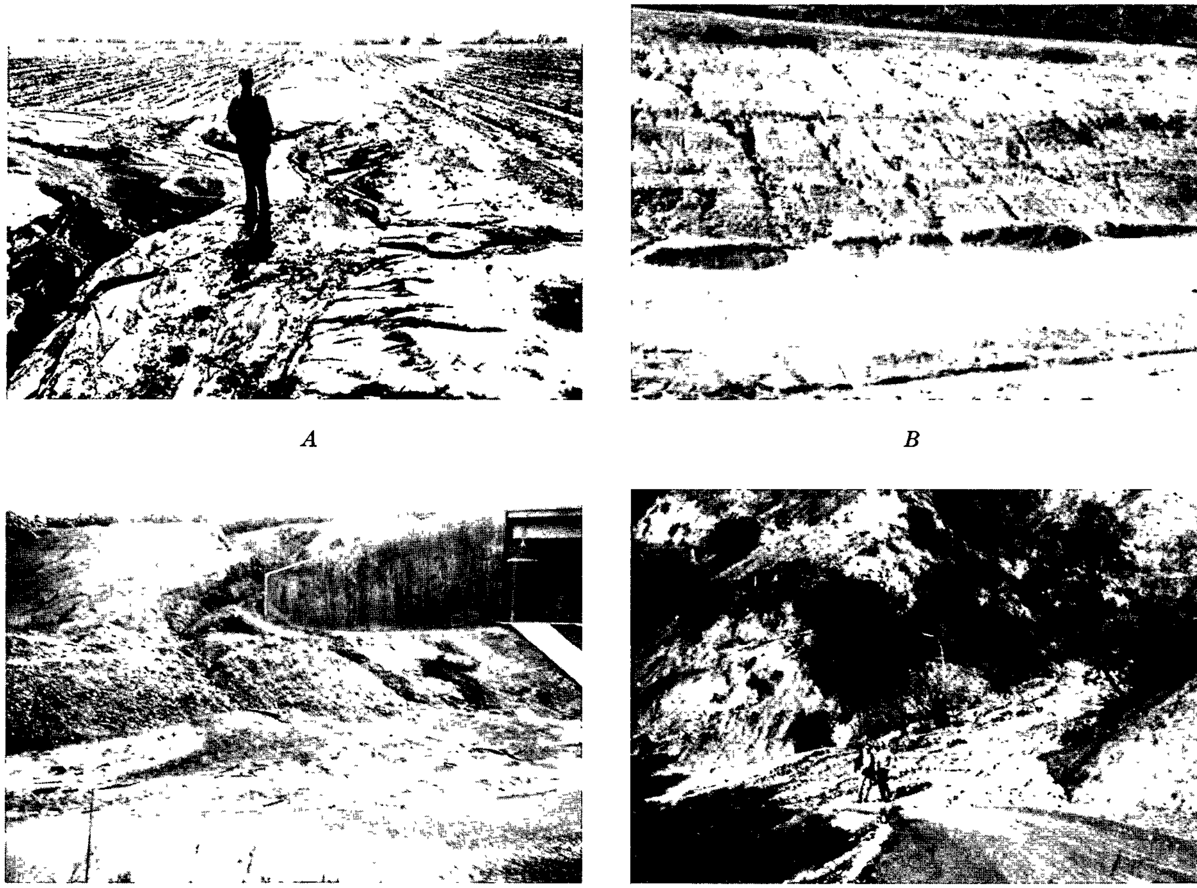
> When water flows over unconsolidated sediment at high enough velocities, some sediment particles are removed from the bed. Of those that are lifted or started into motion, some fall back to the bed but some are carried upward. Even though the number that move upward is only a small fraction of the 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 17*B* 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



A

B

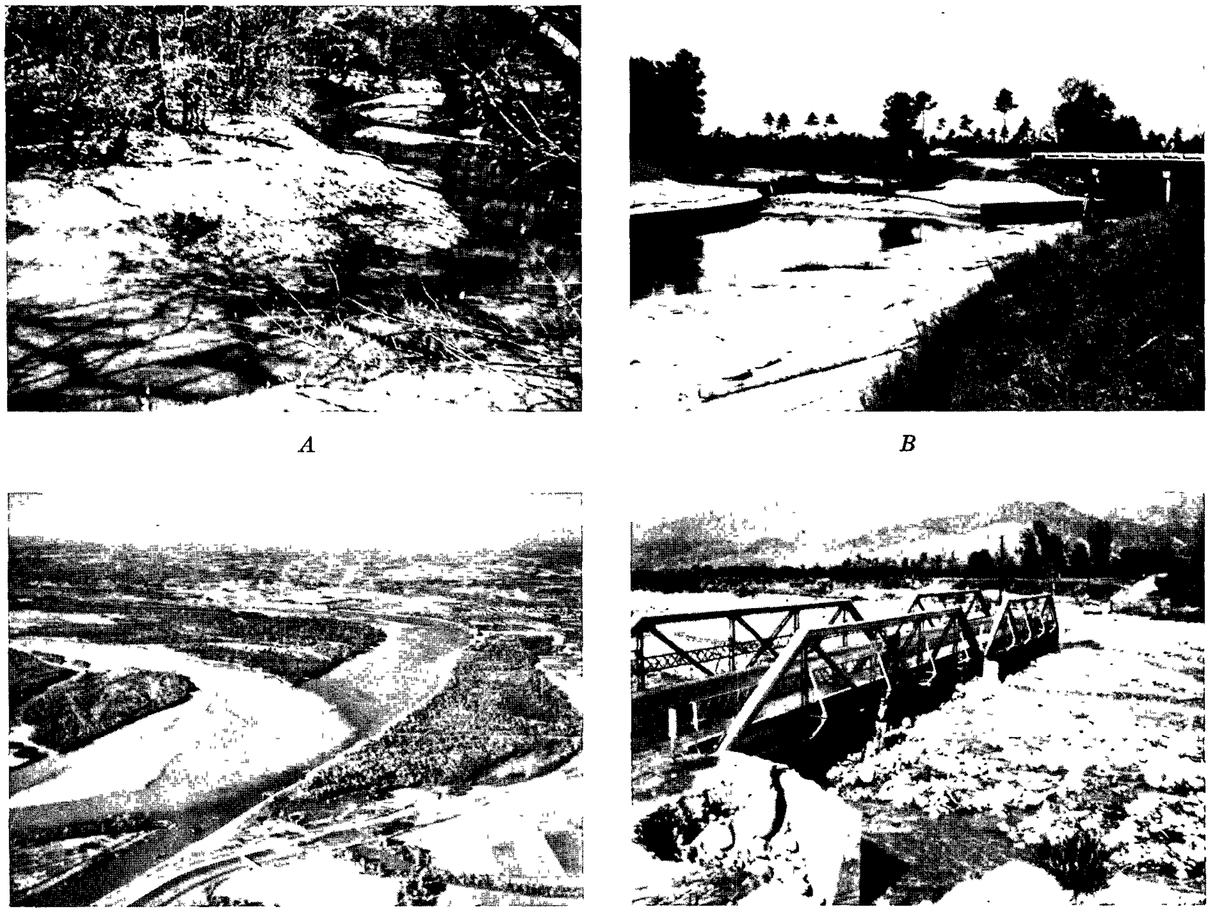
C

D

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.



A B

C D

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).

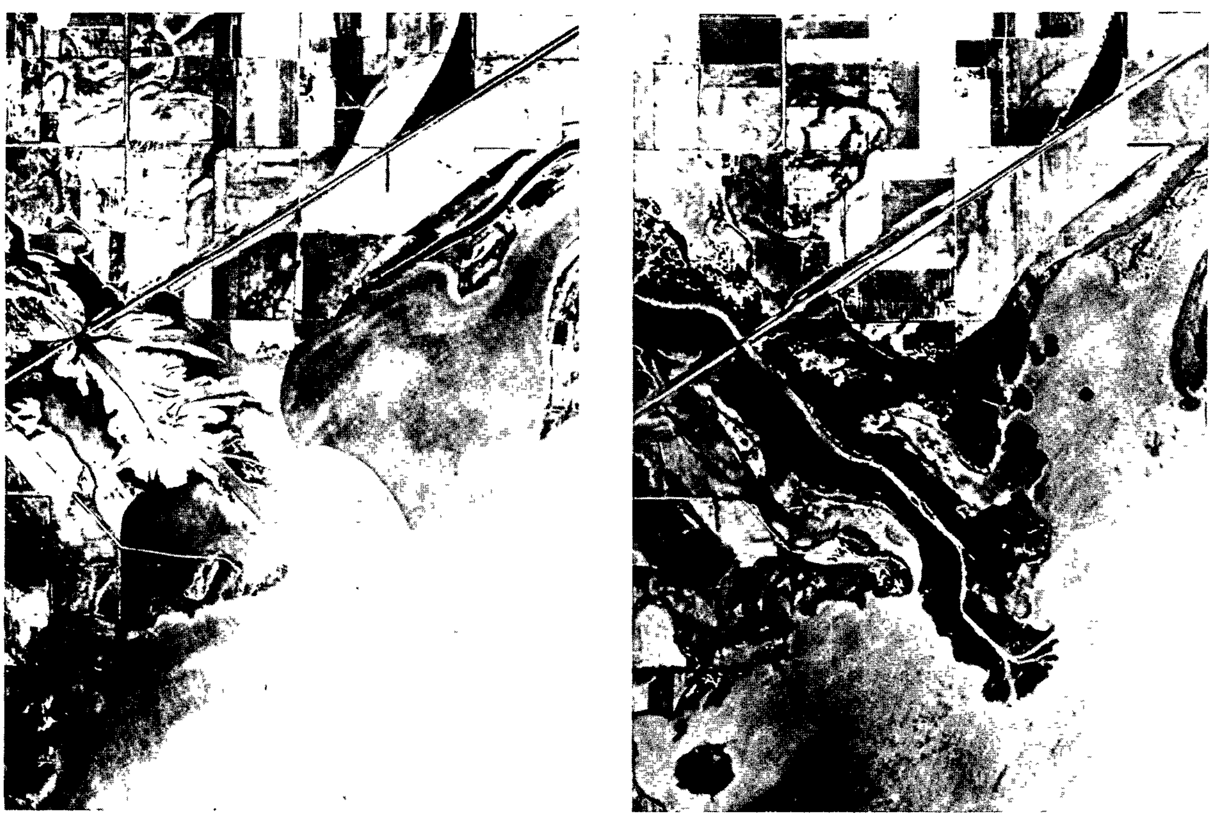
Reservoir deposition

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-

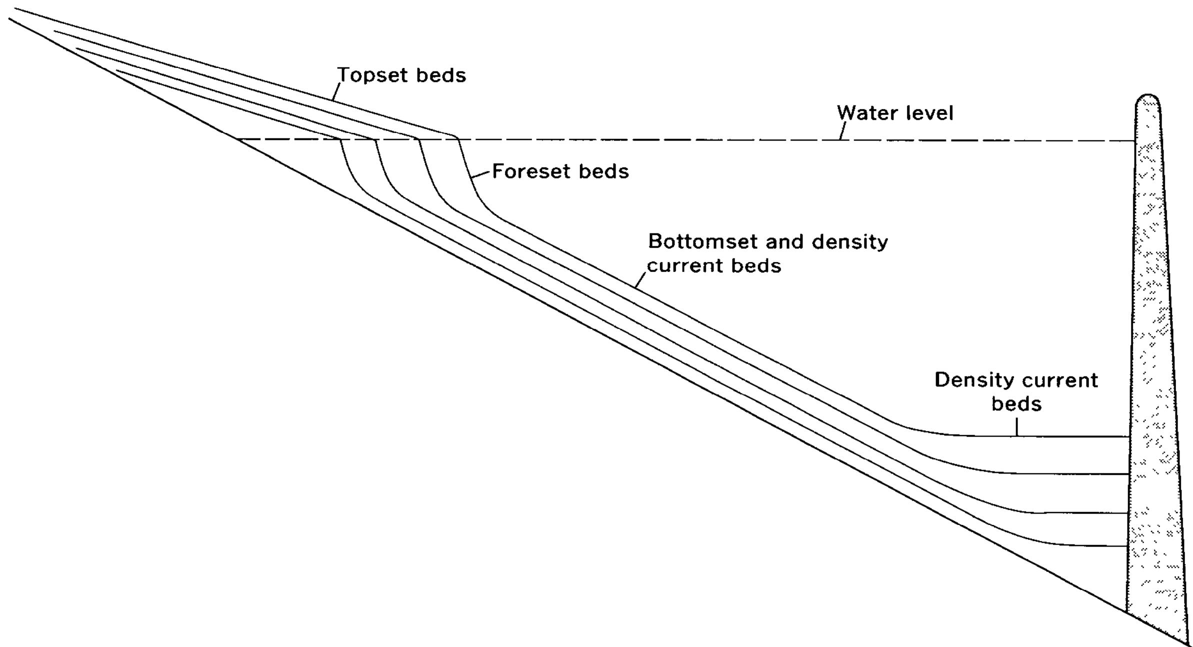


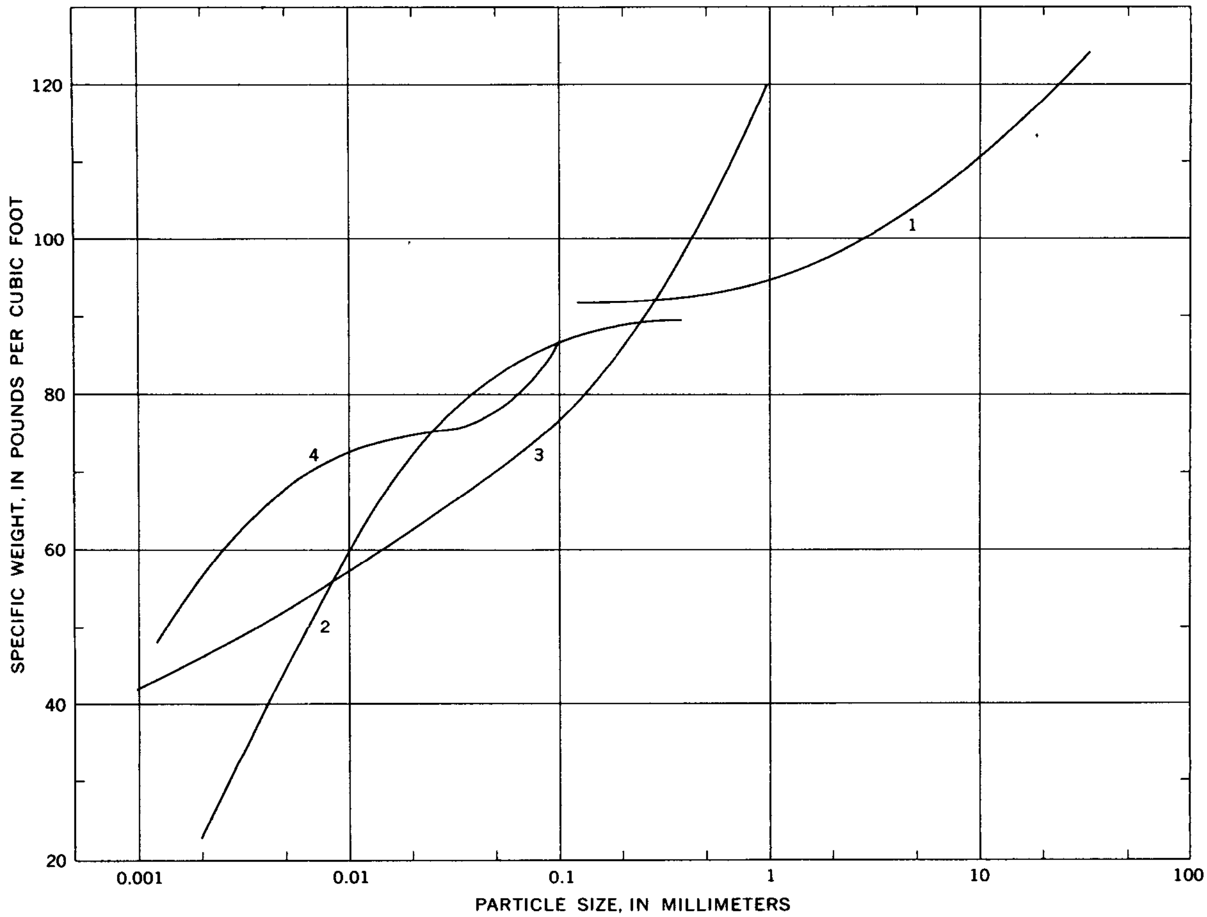
Figure 19.—Longitudinal cross section through a reservoir operating at constant water level. Various types of deposits are shown. Modified from Lane (1953).

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 represented 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.

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

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)
2. Laboratory studies by Parker Trask (U.S. Inter-Agency Report 9, 1943, p. 30)
3. Countrywide (Hembree and others, 1952)
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): Dissolved solids	Average load (tons per sq mi per year): Sediment	Total denudation (inches per 1,000 years)
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

ponent of the eroded slope allows a slightly different system of cross grading and as larger channels pirate or rob smaller ones.

In consideration of a whole drainage basin, Horton (1945) was among the first to recognize the relationship of stream length and stream number to stream order. Stream order is a measure of stream position in the net with respect to its upstream collection. A first-order stream has no tributaries, a second-order stream has only first-order tributaries, a third-order stream has only first- and second-order tributaries, and so forth. Also, the longest tributary from the stream segment of the largest order is extended headward to the drainage area of all streams draining to a site on the stream of the given order. Horton also introduced the term "bifurcation ratio" to express the ratio of the number of streams in a basin of any given order to the number of the next lower order. This ratio tends to equal about 3.5 for many basins in the United States, especially when considering only stream nets shown on maps at a scale of 1:24,000.

In a study of hydrographs from small basins in Pennsylvania, McSparran (1968) defined several drainage-basin characteristics as follows:

1. Area, A, as the square miles of area enclosed by the water divide.
2. Length, L_s, as the distance in miles along the stream to the most remote point on the divide.
3. Slope, S, as the geometric average slope of the profile taken along the stream used to determine L_s.
4. Drainage density, D_d, as the ratio of the total length of all streams in the basin (from USGS 1:24,000-scale maps) to the drainage area.
5. Basin shape factor, F, as the ratio of the length to the remote point, L_s, to the diameter of a circle with an area equal to the drainage area.

Generally basin length, L, is simply defined as the maximum distance from the basin mouth to the water divide, and basin shape factor and slope are defined using L instead of L_s. Schumm (1954) successfully related mean annual sediment loss for a variety of small drainage basins in the Colorado Plateaus province to a basin-relief ratio defined as the ratio between total basin relief and basin length, L. Position along the curve indicates the relative resistance of a given basin to sediment erosion.

Mass wasting

Mass wasting, or the gravitative transfer of material toward and into the streams, has some degree of importance. Too often only the precipitous or very notable types are recognized. Sharpe's classification (1938) of mass-wasting types has come into general usage, and it is sufficient to quote his classes and their definitions directly from Thornbury (1954, p. 45–46).

Slow-flowage types:

Creep: The slow movement downslope of soil and rock debris which is usually not perceptible except through extended observation.

Soil creep: Downslope movement of soil.

Talus creep: Downslope movement of talus or scree.

Rock creep: Downslope movement of individual rock blocks.

Rock-glacier creep: Downslope movement of tongues of rock waste.

Solifluction: The slow downslope flowing of masses of rock debris which are saturated with water and not confined to definite channels.

Rapid-flowage types:

Earthflow: The movement of water-saturated clayey or silty earth material down the low-angle terraces or hillsides.

Mudflow: Slow to very rapid movement of water-saturated rock debris down definite channels.

Debris avalanche: A flowing slide of rock debris in narrow tracks down steep slopes.

Landslides: Those types of movements that are perceptible and involve relatively dry masses of earth debris.

Slump: The downward slipping of one or several units of rock debris, usually with a backward rotation with respect to the slope over which movement takes place.

Debris slide: The rapid rolling or sliding of unconsolidated earth debris without backward rotation of the mass.

Debris fall: The nearly free fall of earth debris from a vertical or overhanging face.

Rockslide: The sliding or falling of individual rock masses down bedding, joint, or fault surfaces.

Rockfall: The free falling of rock blocks over any steep slope.

Subsidence: Downward displacement of surficial earth material without a free surface and horizontal displacement.

Thornbury further states,

The conditions which favor rapid mass wasting were divided by Sharpe (1938) into passive and activating or initiating causes. **Passive causes** include: (1) **lithologic factors,** unconsolidated or weak materials or those which become slippery and act as lubricants when wet, (2) **stratigraphic factors,** laminated or thinly bedded rock and alternating weak and strong or permeable and impermeable beds, (3) **structural factors,** closely spaced joints, faults, crush zones, shear and foliation planes, and steeply dipping beds, (4) **topographic factors,** steep slopes or vertical cliffs, (5) **climatic factors,** large diurnal and annual range of temperature with high frequency or freeze and thaw, abundant precipitation, and torrential rains, and (6) **organic factors,** scarcity of vegetation. **Activating causes** are: removal of support through natural or artificial means, oversteepening of slopes by running water, and overloading through water saturation or by artificial fills.

The reader can recognize from these descriptions that streams can be altered with respect to width, slope, and sediment load by one or more of the many forms of mass wasting. The mudflow, for example, has been treated by Croft (1967) as a problem in public welfare because of its notable occurrence in the form of a "catastrophic event." These can occur on steep-sloped streams draining areas where vegetation and soil have been damaged on a significant part of the drainage basin. Such debris floods are often of short duration, frequently an hour or less, and carry very heavy concentrations of sediment, sometimes with boulders ranging up to several tons in size. Croft (p. 9) reports an hypothesis for the movement of boulders as follows:

While the debris flow is confined to narrow canyon walls, the boulders are almost completely submerged in the semifluid concretelike matrix with a density of about two. The push exerted downslope by the mass and the ball-bearing effect of smaller rocks are important factors in forward motion. An example of movement by pushing and rolling is the 8-ton boulder at the forward end of the Kay Creek mud-rock flood of 1930. This boulder moved about a quarter mile from the canyon mouth across slopes averaging 8.3 percent.

Channel properties

At a given time, the drainage network is a highly organized complex system of physical and hydraulic features which route excess water and weathered products from higher to lower elevation. At a given location in a channel, the tangential stress of flow on the channel boundary is equal and opposite to the resistance exerted by the bed. The transmittal of this shearing stress or exchange of momentum from layer to layer in the flow causes a gradient in the flow velocity. With respect to the energy involved, the slope of the water surface is a direct measure of the energy exchange where there is no velocity change at a point (steady flow), and where there is no change in velocity with distance along the channel (uniform flow). The ultimate fate of the potential energy derived from movement of the flow along the slope is conversion to heat.

With the fact in mind that most of the energy dissipation in open channels is proportional to the square of the flow velocity, Leopold, Wolman, and Miller (1964, p. 162) suggest the possibility of three types of resistance. The first type is skin resistance, caused by the roughness that is in turn determined by the size and character of the material in the bed and banks. For a given roughness, the amount of resistance varies with the square of the flow velocity. The second type is internal distortion resistance, caused by boundary features such as bank protuberances, bends, bars, or individual boulders that set up eddies and secondary circulations. Resistance from these features is also proportional to the square of the mean flow velocity. The third type is spill resistance, where energy is dissipated by local waves and turbulence caused when a sudden reduction in velocity is imposed. In a natural stream these individual resistance types cannot be measured; the total dissipation, however, is indicated by the longitudinal profile of the stream.

Hack (1957) indicates that the longitudinal profile of a stream may be controlled by several factors that are related to both the physical and the chemical properties of the bedrock. Therefore, the sediments found in streams with a given bedrock and similar climate and vegetation are likely to have unique size characteristics at different points along the channel. Hence, stream slopes are expected to be similar for geologically similar areas. Figure 21 from Hack shows how the stream slope changes along its length for several areas of similar bedrock. Such definitive slope patterns would be less distinguishable in larger basins which have more complicated geology, climate, and vegetative controls.

In many streams, vegetation such as grass, weeds, willows, and trees may affect the channels' resistance to flow, especially in the part of the channel above the "normal" flow. Often a high flow will remove, partly remove, or bury the lower types of such vegetation; this removal or burial causes considerable change in resistance during the period of the runoff event.

Omitting vegetation, channel resistance to flow is largely a function of the sizes and shapes of grains or particles, the microfeatures, and the larger boundary or macrofeatures. A bed of large irregular-shaped particles will offer more resistance than a sand-gravel complex. Figure 22 gives the size distribution of bed material for several streams at gaging stations. These distributions represent sizes found for most streams. Note that distributions to the left of a median size (50 percent) of about 1 mm would be called sand-bed streams. The resistance to flow for the different bed forms for sand-bed streams has been discussed on page 16. The distributions with respect to some of the streams plotted in figure 22 also indicate that the particle size of bed material tends to become finer in the downstream direction. Even in the 1,000-mile reach of the Mississippi River between Cairo, Ill., and New Orleans, La., the median size decreases from about 0.65 mm to about 0.20 mm.

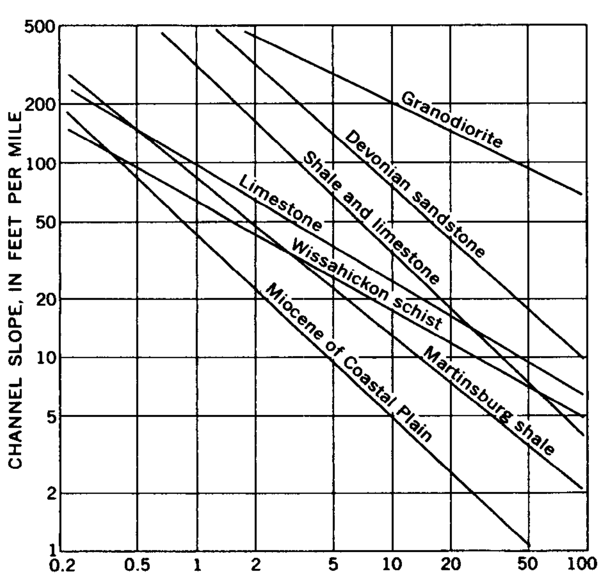
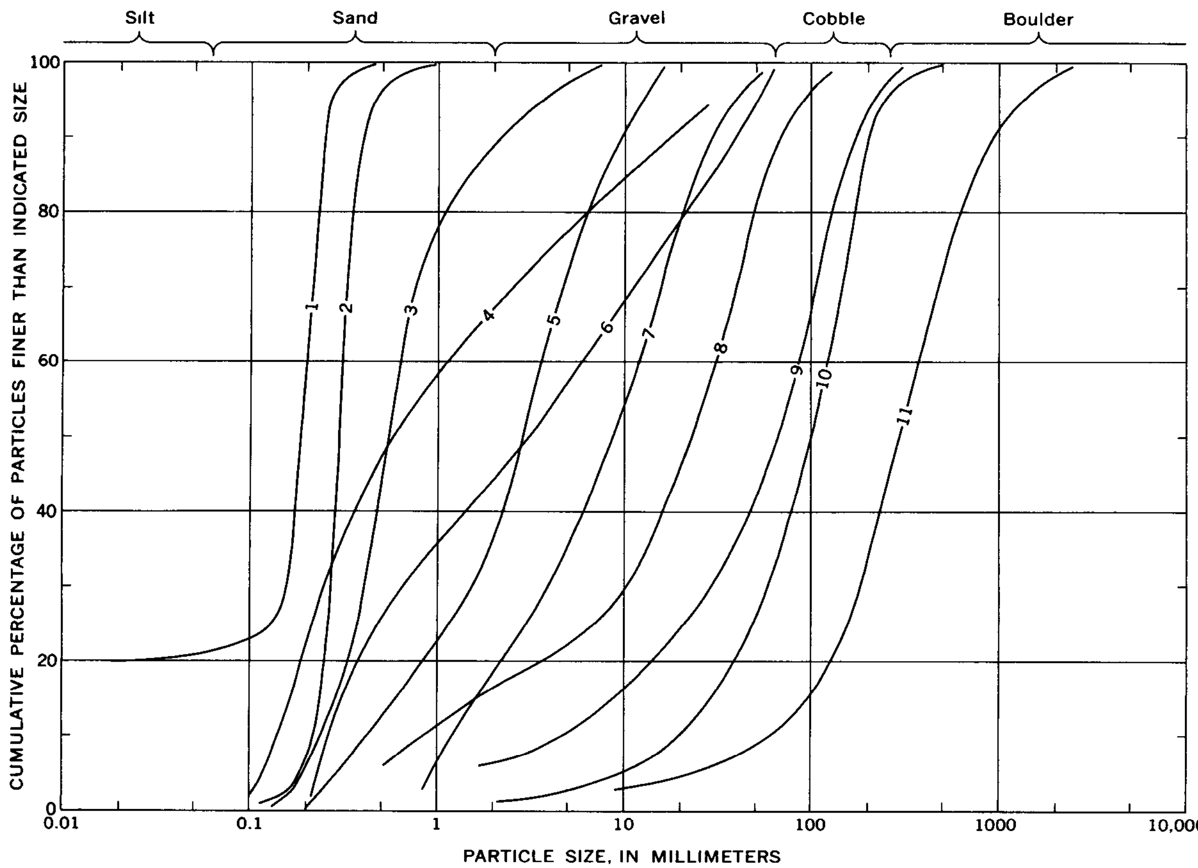


Figure 21.—Average relation between channel slope and stream length for seven geologically different areas in Maryland and Virginia. From Hack (1957, p. 88).

In addition to the bed forms and other macrofeatures already described, it is well to note that sand-bed streams may form large moving bars or sand waves. Carey and Keller (1957) describe sand waves in the Mississippi River as much as 10 meters high and up to 3 km long, on which smaller waves or dunes were noted. Alternate bar formation has also been observed in laboratory flume experiments (Simons and Richardson, 1966). Erosion on the streambank opposite alternate bars may be a factor in the development of stream meanders.

In streams where gravel-sized material or larger is present on the bed, the development of pools and riffles is common, especially in the smaller streams. The spacing of riffles in both straight and meandering channels appears to suggest that the same wave phenomenon that creates the meander is also operating in the straight channel. Riffles in rivers are of lobate shape extending alternately from the banks so that the low-water flow bends around the nose of each riffle. The bends cause a sinuous course even when the stream banks are rather straight.

Alluvial streams characteristically tend to meander; that is, they develop a series of rather symmetrical alternate bends that may grow in lateral extent and at the same time migrate downstream. Among the many who have found empirical relations between such variables as meander length, meander-belt width, channel width, and radius of curvature, Jefferson (1902) was one of the first to recognize specific meander characteristics. Leopold, Wolman, and Miller (1964, p. 298) in a study of stream meanders on 50 rivers of different sizes and from



1 -- Mississippi River at Head of Passes, La.
2 -- Mississippi River at Cairo, Ill.
3 -- Missouri River at Omaha, Nebr.
4 -- Republican River at Clay Center, Kans.
5 -- South Platte River at South Platte, Colo.
6 -- Pembina River at Walhalla, N. Dak.
7 -- Seneca Creek near Rockville, Md.
8 -- Brandywine Creek at Lenape, Pa.
9 -- Brandywine Creek at Cornog, Pa.
10 -- Yellowstone River at Billings, Mont.
11 -- W. Fork Rock Creek near Red Lodge, Mont.

Figure 22.—Particle-size distribution of streambed material typical of indicated streams in the United States.

different physiographic provinces found that the ratio of the radius of curvature to stream width averaged 2.7 and that two-thirds of the values were in the range 1.5 to 4.3. If the meander length (wavelength) is about 10 times the stream width, then the radius of curvature is about one-fourth of the meander length.

The highest velocity of flow in several cross sections around a meander is usually near the concave bank a bit downstream from the axis of the bend. The velocity in a meander crossover is usually somewhat higher on the side of the concave bank upstream. A generalized diagram of the velocity distribution at five cross sections in half a wavelength is shown in figure 23. These velocity patterns in the meander system suggest that the maximum erosion of the concave bank should occur just downstream of the axis of the bend. Friedkin (1945) noted that sand eroded from a concave bank in a laboratory "river" was generally deposited on a point bar downstream on the same side of the channel. This would be expected because the superelevation of the flow toward the concave bank would in turn cause a sidewise current on the streambed from the outside to the inside of the bend. This is suggested to be part of the mechanism of point-bar building and maintenance. The concentration of suspended sediment should be nearly uniform across the section slightly downstream from the crossover (section 1, fig. 23) between the bends because there should be no sidewise current at this location. As the flow moves into and somewhat past the center of the bend (section 3, fig. 23), the intensity of the crosscurrent increases toward the concave bank on the stream surface and toward the convex bank on the streambed. The sidewise current along the bed carries the heavier concentrations and larger particles from the deeper part of the section toward the shallower part near the convex bank.

Experiments with models at the Waterways Experiment Station (Lipscomb, 1952) show that the size of bends (meander length and amplitude) may become smaller with a decrease in flood discharges, the slope, or the angle of entrance to the bend. Moreover, the experiments show that the more erodible are the banks, the wider and shallower will be the crossings between the bends to transport the greater load of sediment from the eroding banks. Because of the fact that the maintenance of channel cross sections and the movement of meanders must be accompanied by the movement of sediment, Benson and Thomas (1966) suggested that the dominant discharge with respect to meanders be defined as that discharge which over a long time period transports the most sediment. Though the highest sediment rates generally occur over a rather large range of flow rates, they found the dominant discharge defined in this manner to be much less than the bankfull stage discharge.

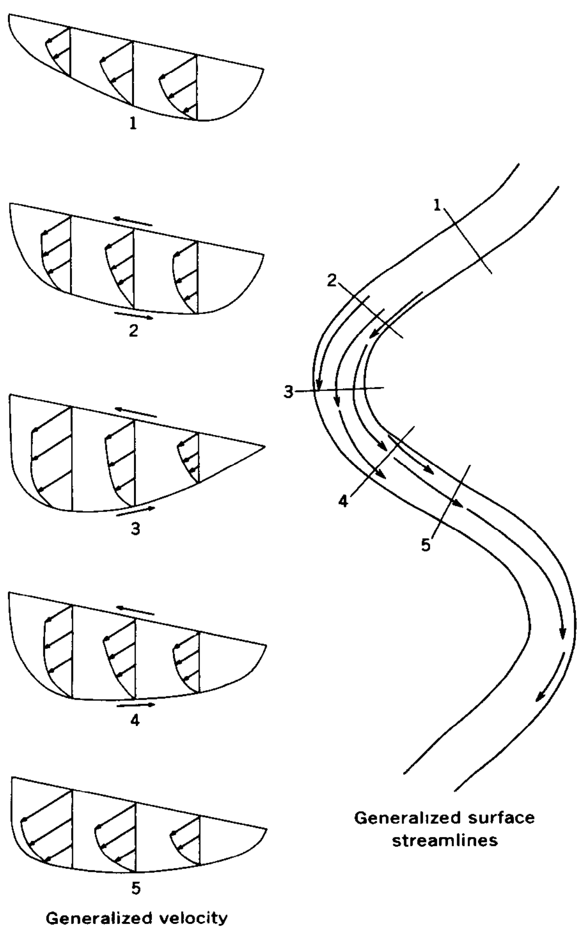


Figure 23.—Diagram of cross-sectional flow distribution in a meander. Note arrows indicating crosscurrents in sections 2, 3, and 4. Modified from Leopold, Wolman, and Miller (1964, p. 300).

The mechanics of meander and stream movement over a flood plain suggests that several features of sediment erosion and deposition may be observed. Some are more noticeable than others on a particular stream, depending on its sediment load and whether or not it is aggrading or degrading. Leopold, Wolman, and Miller (1964, p. 317) list the following features typical of the flood plain:

1. The river channel.
2. Oxbows or oxbow lakes, representing the cutoff portion of meander bends.
3. Point bars, loci of deposition on the convex side of river curves.
4. Meander scrolls, depressions and rises on the convex side of bends formed as the channel migrated laterally downvalley and toward the concave bank.
5. Sloughs, areas of dead water, formed both in meander-scroll depressions and along the valley walls as floodflows move directly downvalley, scouring adjacent to the valley walls.
6. Natural levees, raised berms or crests above the flood-plain surface adjacent to the channel, usually containing coarser materials deposited as floodflows over the top of the channel banks. These are most frequently found at the concave banks. Where most of the load in transit is fine-grained, natural levees may be absent or nearly imperceptible.
7. Backswamp deposits, overbank deposits of finer sediments deposited in slack water ponded between the natural levees and the valley wall or terrace riser.
8. Sand splays, deposits of flood debris usually of coarser sand particles in the form of splays or scattered debris.

In consideration of the geometric and sediment characteristics of the whole stream, it is apparent that a pattern of channel slope and cross section should exist that fits the "dominant" water discharge, the particle-size distribution, and the rate of sediment transport. A diagram (fig. 24) modified from Leopold and Maddock (1953, p. 27) shows how slope, roughness, sediment load, velocity, depth, width, and bed-material size vary with discharge at a station and downstream. Sections A and C represent headwater conditions of low and high flow respectively; B and D represent downstream conditions of low and high flow. Particle size of bed sediment should tend to decrease in the downstream direction and perhaps exhibit a slight increase with increasing flow rate at a site. Note that the indicated change in channel roughness is small in the downstream direction in spite of considerable reduction in skin resistance because of reduced particle size. Most of the reduced resistance from reduced particle size is counteracted by large-scale roughness in the form of increased meanders and (or) sand dunes.

The complexity of stream channels with respect to their shape and the way they may erode, transport, and deposit sediment is indicated in figure 25 (Culbertson and others, 1967). This figure is presented to further indicate the range commonly experienced concerning (*A*) the variability of unvegetated channel width, (*B*) sinuosity, (*C*) bank height, (*D*) natural levees, and (*E*) the modern flood plain.

Economic Aspects

The direct, and most certainly the indirect, economic significance of fluvial sediment problems is usually ignored because many fluvial sediment processes are related to, or are a part of, natural phenomena that often occur in an unnoticed manner. Hence, they are rarely considered for evaluation except when serious consequences can be easily noted and where corrective action is necessary. If the full impact of the erosion of sediment within the river drainage areas, the movement of sediment through stream channels, and the deposition of sediment along streams and in other bodies of water could be evaluated, the subject would be of much greater concern to society.

In a study of damages from sedimentation, Maddock (1969) notes that most information for erosion is presented in terms of loss of plant nutrients, the increased cost of tillage, channel degradation, and loss of land by shore and streambank erosion. For sediment deposits, the counterpart of erosion, most economic information involves maintenance and other costs from infertile material on flood plains, storage loss in reservoirs, channel aggradation, harbors filling, water-supply systems, hydropower turbines, transportation facilities, fish and oyster industries, and wildlife and recreation areas. Because of the subtle nature of sediment damages, this is but a small part of the total damage picture.

Not only may sediment damages go unnoticed, but often they are beyond economic evaluation and have considerable lasting social implication. Maddock states:

> Nevertheless, there are some land areas in the world, such as parts of the Near East and the limestone dolomite region of Yugoslavia, that have become a total loss, economically, during historic times. Nearer to home some agricultural areas of our southeast Coastal Plain have become practically useless through active erosion.

Gottschalk (1965) states:

> Most people have a natural antipathy of "muddy streams." This is particularly evident in fishermen. Aside from the fact that few people care to fish a muddy stream, there is a definite effect of suspended sediment on the size, population, and species of fish in a stream (Ellis, 1936). Suspended sediment affects the light penetration in water and thus reduces the growth of microscopic organisms on which insects and fish feed.

Though only a part of the economic aspects of sedimentation can be presented in terms of dollar damages, a list of several items (table 6) may be helpful to indicate the scope of the problem. As indicated by Ford (1953), it is virtually impossible to separate water damage caused by floods from that caused by a combination of water and sediment. For example, if a flood should cover a crop of wheat in the preharvest stage, the fine sediment in the water will likely impair maturity to a greater extent than if the flood consisted only of clean water. In

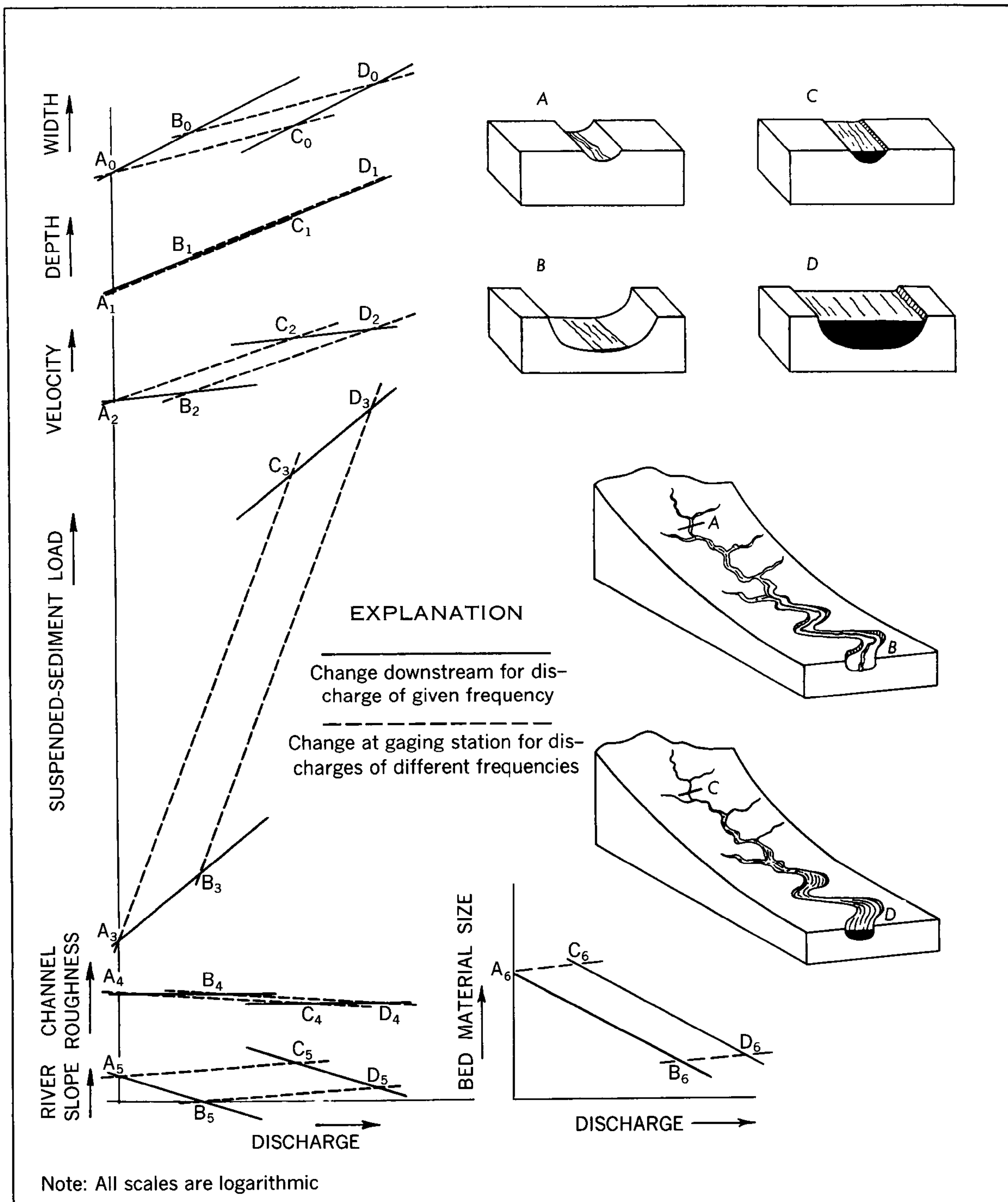


Figure 24.—Average hydraulic geometry of river channels by relations of width, depth, velocity, suspended-sediment load, roughness, slope, and bed-material size to discharge at a station and downstream. Modified from Leopold and Maddock (1953).

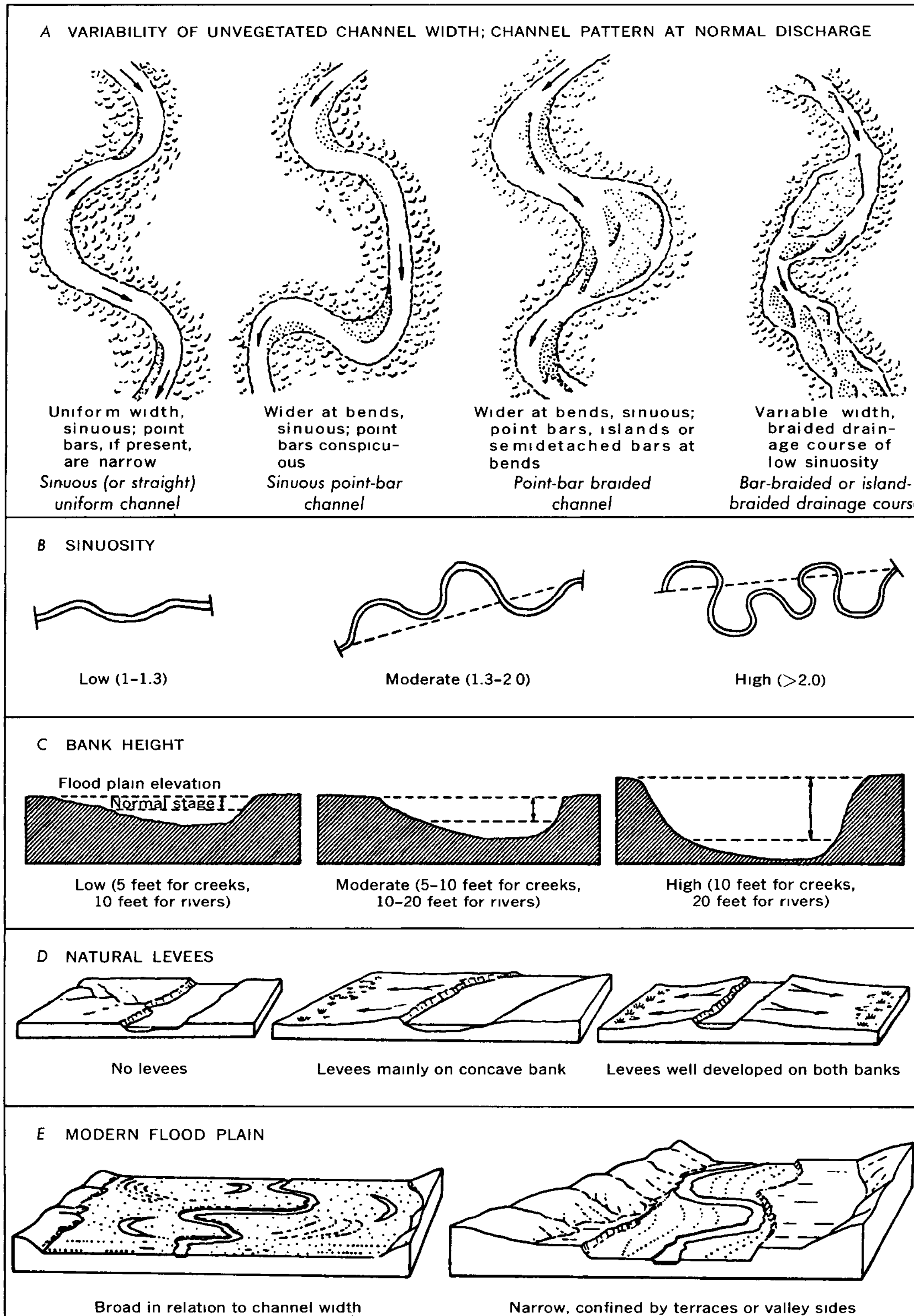


Figure 25.—Complexity of stream channels with respect to channel width, sinuosity, bank height, natural levees, and flood plain. Modified from Culbertson, Young, and Brice (1967, p. 48–49).

flooding of residential property, a large part of the flood damage, especially to household goods, is attributed to sediment in the water. Other types of sediment damage are more easily separated from pure flooding damage. The following broad groups of sediment damages are indicated by Ford: (1) infertile overwash, (2) swamping, (3) filling of reservoirs, (4) damage to water-infiltration facilities, (5) damage to transportation facilities, and (6) damage to drainage and irrigation facilities. Specific items from these groups can be noted in table 6.

Table 6.—Examples of damages from sedimentation

[Most items suggested from Maddock (1969). The damage is not given in dollars of uniform value]

Item	Amount	Basic reference
1. Increased crop production from use of applicable erosion control programs.	An average of $2.50 per acre of all cropland; many examples over $9.00.	Leopold and Maddock (1953).
2. Gully destruction of land in Iowa and Missouri.	Capitalized value to society of $603 per acre.	Weinberger (1965).
3. Decline in crop returns from sheet erosion on Austin clay soil in Texas.	Cumulated loss of $252 per acre as compared to uneroded areas.	Smith, Henderson, Cook, Adams, and Thompson (1967).
4. Infertile overwash, impairment of drainage, channel aggradation, flood-plain scour, and bank erosion.	$50,000,000 annually in United States based on survey of 34 basins representing one-eighth of land area.	Brown (1948).
5. Loss in storage reservoirs used for power, water supply, irrigation, flood control, navigation, recreation, and other multiple purposes.	$50,000,000 annually in United States based on surveys from 600 of the total of 10,000 existing reservoirs.	Brown (1948).
6. Maintenance and impairment of drainage ditches.	$128 for each of the 134,000 sq mi served by such ditches.	Brown (1948).
7. Maintenance of irrigation facilities	About 25 percent of annual total operation and maintenance charge.	Brown (1948).
8. Maintenance of harbors and navigable channels.	$12,000,000 annually (excludes deposits from tidal currents).	Brown (1948).
9. Water purification (excess turbidity)	$5,000,000 annually based on a sample of filter plants.	Brown (1948).
10. Damages during floods; deposits on crops, roads, streets, household effects, and increased flood heights.	$20,000,000 annually as a minimum or about 20 percent of the total flood damages.	Brown (1948).
11. Removal of debris from basins resulting from medium-sized storm in Los Angeles County, 1961.	1,235,000 cu yd at $0.85 (does not include the cost of other extras such as disposal sites).	Ferrell and Barr (1965).
12. Savannah Harbor, Ga	More than $1,000,000 per year to cope with a shoaling rate of 7,000,000 cu yd per year.	Harris (1965).
13. Control of sediment movement at mouth of Columbia River.	Jetty construction $1,969,000 (1895), $9,972,000 (1917), and $6,000,000 (1941).	Lockett (1965).
14. Maintenance of beaches on coastal areas starved for sand by stream controls.	Expensive	Watts (1965).
15. Stabilization of Colorado River below Hoover Dam.	$30,000,000 exclusive of annual maintenance of structures.	Oliver (1965).
16. Reservoir space allotted to sediment storage for four dams on the middle Rio Grande.	$35,000,000 as a part of total cost of dams. Other "sediment" costs of projects not included.	Maddock (1969).
17. Channel erosion in Five Mile Creek near Riverton, Wyo., from effluent of Riverton irrigation project.	$400,000 plus $4,000 maintenance per year	Maddock (1960).
18. Erosion and transport from urban construction of about 5,000,000 acres in United States (mostly urbanization).	Depends on water and land use within and below construction sites.	Guy (1965), Wolman (1964).
19. Erosion and transport from rural cropland areas in United States since settlement.	Forced abandonment of crop production on 35,000,000 acreas.	U.S. Department Agriculture, Agricultural Research Service. (1965).
20. Estimated annual total erosion and sediment problems in United States.	$1,000,000,000	Moore and Smith (1968).

Data needs and program objectives

Data needs

No matter how precise the theoretical prediction of sedimentation processes becomes, it is inevitable that man's activities will increasingly cause the many variables to change relative to their effect on fluvial sediment. There will, therefore, be an increasing need for direct or indirect measurement of fluvial sediment movement and its characteristics to provide data for prediction of the kind and magnitude of sediment problems or to verify the usefulness of a given control measure.

Because of the changing effects of the environment on fluvial sediment, caused mostly by man's activities and the rapid advances in technology, it seems useless to list the many specific kinds of sediment problems we face today. Instead, it is desirable to list only the general areas of concern where many kinds of sediment problems have already occurred and where they may occur in the future.

Water utilization

Water-quality goals and objectives with respect to sediment are being set up with a view to specific domestic, industrial, recreational, and other water uses. Such goals should logically be subject to change as the requirements of use change. Esthetically, for example, a stream should be managed so that it will be more free of sediment when the use is changed from a "private" farming area to a park for public use. Thus, a knowledge of fluvial sediment conditions is needed to help establish criteria for water-quality standards and goals to aid in many aspects of water utilization.

It is difficult to assess the significance of turbidity or sediment concentration in water because of the many simultaneous interactions of detrimental and beneficial effects. Swimming and most recreational uses require nearly sediment-free water; on the other hand, turbid water will reduce or eliminate objectionable algal growth. Sediment is a problem at water-treatment plants because it requires an effort for its removal from the water and its disposal and yet some fine sediment is often desirable in order to effectively remove some organic and inorganic substances in the treatment process. Therefore, considerable monitoring is evidently needed, either in the form of daily or more frequent suspended-sediment measurements or perhaps in the form of a continuous assessment of turbidity as a hydrologic measurement. If turbidity measurement is accomplished, then additional conventional sediment measurements, at least on a periodic basis, will be required in most instances for effective evaluation with respect to water utilization.

Sorption and pollution concentration

The significance of sediment as a sorbing and concentrating agent of pollutants is not well understood with respect to many materials such as organics, pesticides, nutrients, and radionuclides. The organics associated with sediment are highly variable in quantity and tend to interact with many kinds of pollutants in a very complex manner. Because of the complex interaction with sediment, pollutant transport characteristics in streams must necessarily also be very complex. The relatively inert inorganic sediments are not so highly interactive with many pollutants, but they are known to be very important in some instances—two substances which readily affix themselves on sediment are the radionuclide cesium-137 from military weapons and phosphorus from water-treatment plants.

Variation of geomorphological settings

Much of the fluvial-sediment data in the past has been obtained on streams representing large areas of quite diverse environment. It is impossible to obtain data for all streams that have small drainage areas, but it should be possible to greatly increase knowledge concerning the environment-sediment relationship by careful selection of some representative basins for detailed study. If it is impractical to obtain detailed sediment information, it may be possible to use a systematic method of periodic sampling for a large number of basins for which the so-called "rating curve" of suspended-sediment concentration versus water discharge will serve as an empirical guide to environmental effects.

Logically, the reconnaissance type of data program should precede either the periodic or detailed study.

Work concerning the shape of alluvial channels and the erosion and deposition in streams in relation to sediment type and physical characteristics has only been started (Schumm, 1960, 1961). These early studies indicate that the silt-clay content in the channel and the banks affects the width-depth ratio of the stream. A channel composed of fine highly cohesive sediment may have new deposition of a durable nature on the banks as well as the channel floor. Rapid growth of vegetation in these fine sediments may aid such deposition, but it is not necessarily the initial cause of aggradation. If degradation occurs in the fine sediments, it is usually by upstream migration of headcuts. In contrast, a channel containing mostly sands has no deposition of durable deposits in the streambed and little or no "plastering" of fines on the banks. Vegetation is usually sparse on these poorly cohesive, highly mobile sediments. Bank caving is usually more active for the sand-bed stream than for the fine-sediment stream.

Leopold, Emmett, and Myrick (1966) measured the amount of sediment derived from different erosion processes in various physiographic positions in several ephemeral washes draining areas ranging from a few acres to 5 sq mi. The results showed that mass movement, gully-head extension, and channel enlargement are small contributors of sediment compared with sheet erosion on unrilled slopes.

Urban growth

Urban growth has several fluvial-sediment implications. In the construction areas, protective vegetation and topsoils are removed, and drainage areas, slopes, and channels are altered so that the environmental conditions are extremely dynamic with respect to area and time. After construction is complete, the surface erosional pattern may return to a condition somewhat better or worse than for the previous rural setting, but channel erosion will likely be accelerated because of the increased rate and amount of runoff resulting from increased imperviousness in the drainage basin. Although the total area involved with urban growth is small relative to the rural setting, it is worthy of considerable attention because of the dramatic increase in the intensity of sediment erosion, transportation, and deposition in comparison with the rural areas. Urban growth areas are representative of extreme sediment variation with time as well as space and therefore require intensive and detailed study.

Transport and deposition

Sediment transport and deposition processes form the connecting links between the initiation of movement by erosion and the resting place prior to consolidation. Fine-sediment transport occurs when particles finer than most found in the streambed are moved by small fluid forces in nearly continuous suspension. Coarse-sediment transport, on the other hand, occurs when those particles found abundantly in the streambed are moved intermittently by suspension and as bed load. The quantity of fines in the flow at a stream site depends on the release of these fines by erosion and their routing with the flow, whereas the quantity of coarse sediment moved depends on the availability of the specific sizes from the basin to maintain the stream boundary and the energy of the streamflow. Furthermore, the fine sediment tends to disperse with the fluid throughout the stream cross section, whereas the coarse sediment moves mostly near the bed of the stream and at a nonuniform rate across the width of the stream.

Channel aggradation or degradation will occur in a reach of a stream when the transport capacity of the flow does not match the supply of coarse sediment of specific sizes coming into the reach. Deposition problems may occur at any point in the flow system, beginning near areas of excessive erosion and continuing in manmade channels, in natural channels, in ponds and reservoirs, in estuaries, and on beaches. As indicated in several of the examples listed in table 6, the basic problem in connection with deposition is that it usually consists of an accumulation of unwanted material at a location desired for water storage or movement.

One important example relative to transport and deposition data needs concerns scour and fill with respect to structures in channels, particularly highway bridges. Prediction of scour or fill from hydraulic theory and empirical equations has proven uncertain, and hence, there is a great need for case histories to form a base for making better predictions. Culbertson, Young, and Brice (1967) indicate that scour and fill problems may be the result of (1) an increase in stream discharge, (2) an increase or decrease in sediment load relative to water discharge, (3) a change in local base level of the body of water into which the channel flows, (4) a change in channel slope, (5) a lateral shift or redirection of the channel, (6) a downstream progression of a sediment or debris wave, and (7) obstacles or constrictions in the path of flow. Their suggestions for the preparation of a case history on scour and fill include the assembly of such information as (1) photographs and maps, (2) aspects of construction and maintenance of the structure, (3) the morphological properties of the stream, (4) flood history, (5) cross-section and slope surveys, (6) velocity distributions for normal and high flows, (7) bed- and suspended-sediment discharge rates including particle-size distributions, and (8) the characteristics of bed forms including the depth of scour around piers and abutments.

Program objectives

In consideration of the many general problem areas in sedimentation, it is aximatic that program objectives, if they are quite specific, would have to be very flexible to meet the ever-changing set of problems. Unusually, however, a set of general objectives that are more stable can form the basis of the dynamic detailed objectives. An example of a set of these objectives was presented by R. B. Vice at Albuquerque in April 1967:

1. Develop and maintain a national network of sediment-measuring stations to provide unbiased comprehensive information about sediment movement in streams.
2. Study and describe sedimentation in specific priority areas so that water managers will have at hand essential information for choosing between alternatives.
3. Expand research studies in sedimentation to disclose and describe process relationships between water, sediment, and the environment.

Network and aerial coverage

Exclusive of special and local sediment problems, the World Meteorological Organization's "Guide to Hydrometeorological Practices" suggests a minimum design for a stream-sediment network to include 30 percent and 15 percent of the gaging stations in arid and humid regions, respectively. The extent of coverage for a specific budget is directly related to the unit cost, which in turn is a function of the size and complexity of the stream system and measurement site as well as a function of the kind and intensity of the sediment-sampling program. Data from sediment networks must provide a basis for the future prediction of events. Therefore, statistics relative to sediment movement and its related environment should include instantaneous and average characteristics as well as the range, variation, and patterns of fluctuations. Whetstone and Schloemer (1967) stress that "the value of data increases with quantity and quality, and therefore data should be systemically preserved." The availability of the electronic computer makes it feasible to reduce and codify data for effective storage and retrieval. The computer also makes possible more sophisticated approaches to hydrologic analysis.

Vice and Swenson (1965) state that a network is an orderly system for acquiring data. They further indicate that the fundamental elements of a network system should include (1) a distribution of stations where repetitive observations can be made that will describe the character and variability in time and space, (2) an evaluation of significant environmental features, (3) the evolvement of improved techniques of data collection, and (4) a continuing program for analysis and interpretation of available data to guide in refinement of the total system.

Present and future benefits in land and water management determine the optimum distribution of sediment data needed for a region. Thus a part of a region in the path of urban development must necessarily receive more intensive coverage than a part of the region set to a minor use. Vice and Swenson (1965) suggest that a beginning network can sometimes be approximated from existing sediment programs that have evolved in response to urgent water and sediment problems. They caution, though, that

> greater effort should be applied to (1) areas of abundant water supply, where large water use can be expected, (2) areas of high sediment variability, where more detail is needed, and (3) areas of high sediment concentration, where sediment is more likely to limit project feasibility.

Wallis and Anderson (1965) in a study of sediment yields from California drainage basins found that man's activities have increased sediment loads by 17 times, and therefore, "a well-designed sedimentation network must be flexible enough to allow for evaluation of the effect of changing land use."

Though the prediction of future events is probably the most important purpose to be served by a sediment network, the basic sediment network should often be supplemented by additional programs. These may be programed to provide detailed information on the location of erosion areas and the relative amount of the eroded material that is deposited at different locations within the basin. Special studies may also be required (1) to evaluate erosion-control programs applied to problem areas, (2) to determine the effects of interbasin water diversions, (3) to monitor sediment transport within and from areas of urban development, and (4) to evaluate the stress on urban channels from increased runoff.

Kinds of site records

The sediment-sampling program at a stream site can be considered to fall into one of three classes. The first is the continuous sediment record, usually called the daily station, in which the amount of sediment as measured by suspended-sediment samples is computed and recorded on a daily basis. A set of suspended-sediment samples should represent the sediment concentration of the stream at the time of the sample, and therefore, the data indicated by the sample must be extended backward and forward in time. The length of time applicable to a given sample depends on the time of the previous and next sample and whether or not there are important changes in stream conditions.

A good program for a daily station, then, requires not only the use of proper equipment to obtain good representative samples but also a very sophisticated set of instructions and judgment with respect to timing of samples. Such a program also depends on the major use of the data. If the problem considers mostly the needs of a water user withdrawing a relatively uniform amount, then the major emphasis should be on the sediment concentration of the flow, and thus the samples would be spaced rather uniformly in time. If the problem concerns the amount or tonnage of sediment moved by the stream, then it may be desirable to sample the low-flow periods once a week or on days of change and to sample two or three times a day during high-flow periods. The thunderstorm type of hydrograph is perhaps the most difficult to sample adequately because of the effects of uneven precipitation in the basin and because of the ever-changing environmental factors, many of which can be related to season of the year and to land use.

The second type of sediment-sampling program can be classified as a partial-record site. This is essentially the same as the daily record except that data are obtained only during selected times of the year based on a predictable period of high flow, or flow greater than a selected rate. The equipment used and the timing of samples for the partial record would be the same as for the daily record.

The third program is the periodic sediment record that may be represented by one of a large variety of sample techniques and timing. Perhaps the most common program would be the collection of samples for a sediment-discharge measurement each time a technician visits the station—once every 2 weeks or once a month, perhaps with more frequent observations during flood periods. These kinds of data provide information for publication of "instantaneous" values of water discharge, sediment concentration, and sediment discharge.

A series of reconnaissance measurements should usually be made prior to the establishment of any of the three programs to obtain comparative information on conditions likely to occur in the future. Even after a program is started, it should be expected that operational adjustments will be required with respect to equipment, sample timing, or even measurement location, especially in areas of changing land use.

Sometimes the requirements of any of these programs may be such that sediment must be measured in terms of total load, in which case it will be necessary either to sample the sediment at a site where it is suspended into the sampling zone by natural or artificial means, or to calculate the amount of the unmeasured sediment. As one would expect, any of the three programs requires a wide range of sampling arrangements determined by climate and drainage-basin characteristics, especially size. The data needs and the operation of a sediment-measuring station on the Missouri River at Kansas City, for example, are vastly different from the needs and operation of a station on a small channel draining a 10-acre basin in an area under urban development.

References

Alger, G. L. and Simons, D. B., 1968, Fall velocity of irregular shaped particles: Am. Soc. Civil Engineers Proc., v. 94, HY3, p. 721–737.

Baver, L. D., 1948, Soil physics [2d ed.]: New York, John Wiley & Sons, 398 p.

Bennett, H. H., 1939, Soil conservation: New York, McGraw-Hill Book Co., 993 p.

Benson, M. A. and Thomas, D. M., 1966, A definition of dominant discharge: Internat. Assoc. Sci. Hydrology Bull., v. 11, no. 2, p. 76–80.

Borland, W. M., 1961, Sediment transport of glacier-fed streams in Alaska: Jour. Geophys. Research, v. 66, no. 10, p. 3347.

Brice, J. C., 1958, Origin of steps on loess-mantled slopes: U.S. Geol. Survey Bull. 1071–C, p. 69–85.

——— 1966, Erosion and deposition in the loess-mantled Great Plains, Medicine Creek Drainage Basin, Nebraska: U.S. Geol. Survey Prof. Paper 352–H, p. 255–339.

Brown, C. B., 1948, Perspective on sedimentation—purpose of conference *in* Proc. of Federal Inter-Agency Sedimentation Conf., Denver, Colo., 1947: U.S. Bur. Reclamation, p. 307.

Brune, G. M., 1953, Trap efficiency of reservoirs: Am. Geophys. Union Trans., v. 34, no. 3, p. 407–418.

Bushnell, T. M., 1944, The story of Indiana soils: Purdue Univ. Agr. Expt. Sta., Spec. Circ. 1, Lafayette, Ind.

Carey, W. C. and Keller, M. D., 1957, Systematic changes in the beds of alluvial rivers: Am. Soc. Civil Engineers Proc., v. 83, paper 1331, 24 p

Colby, B. R., 1961, Effect of depth of flow on discharge of bed material: U.S. Geol. Survey Water-Supply Paper 1498–D, 12 p.

——— 1963, Fluvial sediments—a summary of source, transportation deposition, and measurement of sediment discharge: U.S. Geol. Survey Bull. 1181–A, 47 p.

——— 1964a, Discharge of sands and mean-velocity relationships in sand-bed streams: U.S. Geol. Survey Prof. Paper 462–A, 47 p.

——— 1964b, Scour and fill in sand-bed streams: U.S. Geol. Survey Prof. Paper 462–D, 32 p.

Cook, H. L., 1936, The nature and controlling variables of the water erosion process: Soil Sci. Soc. America Proc., v. 1, p. 487–494.

Corey, A. T., 1949, Influence of shape on the fall velocity of sand grains: Fort Collins, Colo., Colorado State University, M.S. thesis, 102 p.

Croft, A. R., 1967, Rainstorm debris floods: Arizona Univ. Agr. Expt. Sta. Rept. 248, 36 p.

Culbertson, D. M., Young, L. E., and Brice, J. C., 1967, Scour and fill in alluvial channels with particular reference to bridge sites: U.S. Geol. Survey open-file rept., 58 p.

Dawdy, D. R., 1961, Depth-discharge relations of alluvial streams—discontinuous rating curves: U.S. Geol. Survey Water-Supply Paper 1498–C, 16 p.

Dole, R. B., and Stabler, H., 1909, Denudation: U.S. Geol. Survey Water-Supply Paper 234, p. 78–93.

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

Ellis, M. M., 1936, Erosion silt as a factor in aquatic environments: Ecology, v. 17, no. 1, p. 29–42.

Ferrell, W. R. and Barr, W. R., 1965, Criteria and methods for use of check dams in stabilizing channel banks and beds *in* Proc. Federal Inter-Agency Sedimentation Conf., Jackson, Miss., 1963: U.S. Dept. Agriculture, Agr. Research Service Misc. Pub. 970, p. 376–386.

Ford, E. C., 1953, Upstream floodwater damages: Jour. Soil Water Conserv., v. 8, p. 240–246.

Friedkin, J. F., 1945, A laboratory study of the meandering of alluvial rivers: U.S. Waterways Eng. Expt. Sta., 40 p.

Geiger, A. F., 1965, Developing sediment storage requirements for upstream retarding reservoirs *in* Proc. of Federal Inter-Agency Sedimentation Conf., Jackson, Miss., 1963: U.S. Dept. Agriculture, Agr. Research Service Misc. Pub. 970, p. 881–885.

Glymph, L. M., Jr., 1951, Relation of sedimentation to accelerated erosion in the Missouri River Basin: U.S. Dept. Agriculture, SCS–TP–102, p. 6.

Goldich, S. S., 1938, A study in rock weathering: Jour. Geology, v. 46, p. 17–58.

Gottschalk, L. C., 1965, Nature of sediment problems: Am. Soc. Civil Engineers Sedimentation Manual, HY2, p. 259–266.

Gottschalk, L. C. and Jones, V. H., 1955, Valleys and hills, erosion and sedimentation *in* Water: U.S. Dept. Agriculture Yearbook, p. 135–143.

Guy, H. P., 1964, An analysis of some storm-period variables affecting stream sediment transport: U.S. Geol. Survey Prof. Paper 462–E, 46 p.

——— 1965, Residential construction and sedimentation at Kensington, Md., *in* Proc. of Federal Inter-Agency Sedimentation Conf., Jackson, Miss., 1963: U.S. Dept. Agriculture, Agr. Research Service Misc. Pub. 970, p. 30–37.

——— 1969, *Laboratory theory and methods for sediment analysis*: U.S. Geol. Survey Techniques Water-Resources Inv., book 5, chap. C1, 58 p.

Guy, H. P. and Norman, V. W., 1970, Field methods for measurement of fluvial sediment: U.S. Geol. Survey Techniques Water-Resources Inv., book 3, chap. C2.

Guy, H. P. and Simons, D. B., 1964, Dissimilarity between spatial and velocity-weighted sediment concentrations: U.S. Geol. Survey Prof. Paper 475–D, p. 134–137.

Guy, H. P., Simons, D. B., and Richardson, E. V., 1966, Summary of alluvial channel data from flume experiments 1956–61: U.S. Geol. Survey Prof. Paper 462–I, 96 p.

Hack, J. T., 1957, Studies of longitudinal stream profiles in Virginia and Maryland: U.S. Geol. Survey Prof. Paper 294–B, p. 45–94.

Happ, S. C., 1944, Significance of texture and density of alluvial deposits in the Middle Rio Grande Valley: Jour. Sed. Petrology, v. 14, p. 3–19.

Happ, S. C., Rittenhouse, G.. and Dobson, G. C., 1940, Some principles of accelerated stream and valley sedimentation: U.S. Dept. Agriculture Tech. Bull. 695, 133 p.

Harris, J. W., 1965, Means and methods of inducing sediment deposition and removal *in* Proc. of Federal Inter-Agency Sedimentation Conf., Jackson, Miss., 1963: U.S. Dept. Agriculture, Agr. Research Service Misc. Pub. 970, p. 669–674.

Heidel, S. S., 1956, The progressive lag of sediment concentration with flood waves: Am. Geophys. Union Trans., v. 37, no. 1, p. 56.

Hembree, C. H., Colby, B. R., Swenson, H. A., and Davis, J. H., 1952, Sedimentation and chemical quality of water in the Powder River drainage basin, Wyoming and Montana: U.S. Geol. Survey Circ. 170.

Holeman, J. N., 1968, The sediment yield of major rivers of the world: Water Resources Research, v. 4, no. 4, p. 737–747.

Horton, R. E., 1945, Erosional development of streams and their drainage basins, hydrophysical approach to quantitative morphology: Geol. Soc. America Bull., v. 56, p. 275–370.

Hubbell, D. W., 1960, Investigations of some sedimentation characteristics of sand-bed streams, Progress report No. 2: U.S. Geol. Survey open-file rept., 78 p.

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

Ireland, H. A., Sharpe, C. F., and Eargle, D. H., 1939, Principles of gully erosion in the Piedmont of South Carolina: U.S. Dept. Agriculture Tech. Bull. 633, 143 p.

Jefferson, M. S. W., 1902, limiting width of meander belts: Natl. Geo. Mag., v. 13, no. 10, p. 373–384.

Johnson, A. W., 1961, Highway erosion control: Am. Soc. Agr. Engineers Trans., v. 4, no. 1, p. 144–152.

Judson, S., and Ritter, D. F., 1964, Rates of regional denudation in the United States: Jour. Geophys. Research, v. 69, no. 16, p. 3395–3401.

Koelzer, V. A. and Lara, J. M., 1958, Densities and compaction rates of deposited sediment: Am. Soc. Civil Engineers Proc., Paper 1603, v. 84, no. HY2.

Lane, E. W., 1947, Report of the subcommittee on sediment terminology: Am. Geophys. Union Trans., v. 28, no. 6, p. 937.

——— 1953, Some aspects of reservoir sedimentation: Irrig. and Power Jour. Repts. 2 and 3, v. 10, Apr. and July, 1953.

——— 1955, The importance of fluvial morphology in hydraulic engineering: Am. Soc. Civil Engineers Proc., v. 81, no. 475.

Leopold, L. B., Emmett, W. W. and Myrick, R. M., 1966, Channel and hillslope processes in a semiarid area, New Mexico: U.S. Geol. Survey Prof. Paper 352–G, 49 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, 57 p.

Leopold, L. B., Wolman, M. G. and Miller, J. P., 1964, Fluvial processes in geomorphology: San Francisco, Calif., W. H. Freeman and Co., 522 p.

Lipscomb, E. B., 1952, The technique and application of hydraulic model studies involving movable beds: Hydraulics Conf., 5th, Iowa Univ., Iowa City, 1952, Proc., p. 47–65.

Lockett, J. B., 1965, Phenomena affecting improvement of the lower Columbia estuary and entrance *in* Proc. of Federal Inter-Agency Sedimentation Conf., Jackson, Miss., 1963: U.S. Dept. Agriculture, Agr. Research Service Misc. Pub. 970, p. 626–669.

Lustig, L. K. and Busch, R. D., 1967, Sediment transport in Cache Creek drainage basin in the coast ranges west of Sacramento, Calif.: U.S. Geol. Survey Prof. Paper 562–A, 36 p.

Lyon, T. L. and Buckman, H. O., 1943, The nature and properties of soils [4th ed.]: New York, Macmillan Co., 499 p.

Lyons, H. G., 1906, The physiography of the river Nile and its basin: Cairo, Egypt, Survey Dept., 411 p.

Maddock, Thomas, Jr., 1960, Erosion control of Five Mile Creek, Wyoming: Internat. Assoc. Sci. Hydrology, Pub. 53, p. 170–181.

——— 1969, Economic aspects of sedimentation: Am. Soc. Civil Engineers Sedimentation Manual, HY1, p. 191–207.

McSparran, J. E., 1968, Design hydrographs for Pennsylvania watersheds: Am. Soc. Civil Engineers Proc., v. 94, HY4, p. 937–960.

Moore, W. R. and Smith, C. E., 1968, Erosion control in relation to watershed management: Am. Soc. Civil Engineers Proc., v. 94, IR 3, p. 321–331.

Mundorff, J. C., 1966, Sedimentation in Brownell Creek Subwatershed No. 1, Nebraska: U.S. Geol. Survey Water-Supply Paper 1798–C, 49 p.

——— 1968, Fluvial sediment in the drainage area of K–79 Reservoir, Kiowa Creek Basin, Colorado: U.S. Geol. Survey Water-Supply Paper 1798–D, 26 p.

Oliver, P. A., 1965, Some economic considerations in river control work *in* Proc. of Federal Inter-Agency Sedimentation Conf., Jackson, Miss., 1963: U.S. Dept. Agriculture, Agr. Research Service, Misc. Pub. 970, p. 442–449.

Piest, R. F., 1965, The role of the large storm as a sediment contributor *in* Proc. of Federal Inter-Agency Sedimentation Conf., Jackson, Miss., 1963: U.S. Dept. Agriculture, Agr. Research Service, Misc. Pub. 970, p. 98–108.

Piest, R. F., 1970, Sediment sources and sediment yields: Am. Soc. Civil Engineers Sedimentation Manual (in press).

Porterfield, G. and Dunnam, C. A., 1964, Sedimentation of Lake Pillsbury, Lake County, Calif.: U.S Geol. Survey Water-Supply Paper 1619–EE, 46 p

Reiche, Parry, 1950, A survey of weathering processes and products [revised ed.]: Albuquerque, Univ. of New Mexico Press, 95 p.

Roehl, J. W., 1962, Sediment source areas, delivery ratios, and influencing morphological factors: Internat. Assoc. Sci. Hydrology, Pub. 59, p. 202–213.

Russell, M. B., 1957, Physical properties *in* Soil: U.S. Dept. Agriculture Yearbook, p. 31–38.

Sayre, W. W., Guy, H. P. and Chamberlain, A. R., 1963, Uptake and transport of radio-nuclides by stream sediments: U.S. Geol. Survey Prof. Paper 433–A, 33 p.

Schumm, S. A., 1954, The relation of drainage basin relief to sediment loss: Pub. Internat. Assoc. Hydrology, Internat. Union of Geodesy and Geophysics, Tenth General Assembly, Rome, v. 1, p. 216–219.

——— 1960, The shape of alluvial channels in relation to sediment type: U.S. Geol. Survey Prof. Paper 352–B, 13 p.

——— 1961, The effect of sediment characteristics on erosion and deposition in ephemeral-stream channels: U.S. Geol. Survey Prof. Paper 352–C, 39 p.

——— 1963, The disparity between present rates of denudation and orogeny: U.S. Geol. Survey Prof. Paper 454–H, 13 p.

Schumm, S. A. and Hadley, R. F., 1957, Arroyos and the semi-arid cycle of erosion: Am. Jour. Sci., v. 255, p. 161–174.

Sharpe, C. F. S., 1938, Landslides and related phenomena: New York, Columbia Univ. Press, 137 p.

Simons, D. B. and Richardson, E. V., 1962, The effect of bed roughness on depth-discharge relations in alluvial channels: U.S. Geol. Survey Water-Supply Paper 1498–E, 26 p.

——— 1966, Resistance to flow in alluvial channels: U.S. Geol. Survey Prof. Paper 422–J, 61 p.

Simons, D. B., Richardson, E. V. and Haushild, W. L., 1963, Some effects of fine sediment on flow phenomena: U.S. Geol. Survey Water-Supply Paper 1498–G, 47 p.

Simonson, R. W., 1957, What soils are *in* Soil: U.S. Agriculture Yearbook, p. 17–31.

Smith, R. M., Henderson, R. C., Cook, E. D., Adams, J. E., and Thompson, D. O., 1967, Renewal of desurfaced Austin clay: Soil Sci., v. 103, no. 3.

Smith, W. O., Vetter, C. P., Cummings, G. B., and others, 1960, Comprehensive survey of sedimentation in Lake Mead, 1948–49: U.S. Geol. Survey Prof. Paper 295, 254 p.

Spraberry, J. A., 1964, Summary of reservoir sediment deposition surveys made in the United States through 1960: U.S. Dept. Agriculture, ARS, Misc. Pub. 964, 61 p.

Stallings, J. H., 1957, Soil conservation: Englewood Cliffs, N.J., Prentice-Hall, Inc., 575 p.

Stringham, G. E., Simons, D. B. and Guy, H. P., 1969, The behavior of large particles falling in quiescent liquids: U.S. Geol. Survey Prof. Paper 562–C, 36 p.

Thompson, J. R., 1964, Quantitative effect of watershed variables on rate of gully-head advancement: Am. Soc. Agr. Engineers Trans., v. 7, no. 1, p. 54–55.

Thornbury, W. D., 1954, Principles of geomorphology: New York, John Wiley & Sons, Inc., 618 p.

U.S. Department of Agriculture, 1938, Soils and Men: U.S. Dept. Agriculture Yearbook, table 3, p. 996–1001.

U.S. Department of Agriculture—Agricultural Research Service, 1965, Losses in agriculture: U.S. Dept. Agriculture Handb. 291, p. 98–100.

U.S. Department of Agriculture—Soil Conservation Service, 1960, Soil classification, a comprehensive system, 7th approximation, compiled by the Soil Survey Staff: U.S. Dept. Agriculture Soil Conserv. Service, 265 p.

U.S. Inter-Agency Report, 1943, Density of sediments deposited in reservoirs: Minneapolis, Minn., St. Anthony Falls Hydraulic Laboratory, U.S. Inter-Agency Rept. 9, 60 p.

U.S. Inter-Agency Report, 1957, A study of methods used in measurement and analysis of sediment loads in streams: Minneapolis, Minn., St. Anthony Falls Hydraulic Laboratory, U.S. Inter-Agency Rept. 12, 55 p.

Vice, R. B., Guy, H. P. and Ferguson, G. E., 1969, Sediment movement in an area of suburban highway construction, Scott Run Basin, Fairfax County, Virginia, 1961–64: U.S. Geol. Survey Water-Supply Paper 1591–E, 41 p.

Vice, R. B. and Swenson, H. A., 1965, A network design for water quality *in* World Meteorol. Organization and Internat. Assoc. of Sci. Hydrology—Symposium, Design of Hydrol. Networks: Pub. 67, v. 1, p. 325–335.

Wallis, J. R. and Anderson, H. W., 1965, An application of multivariate analysis to sediment network design *in* World Meteorol. Organization and Internat. Assoc. of Sci. Hydrology—Symposium, Design of Hydrol. Networks: Pub. 67, v. 1, p. 357–378.

Watts, G. M., 1965, Sediment discharge to the coast as related to shore processes *in* Proc. of Federal Inter-Agency Sedimentation Conf., Jackson, Miss., 1963: U.S. Dept. Agriculture, Agr. Research Service Misc. Pub. 970, p. 738–747.

Weinberger, M. L., 1965, Loss of income from gullied lands: Jour. Soil and Water Conservation, v. 20, p. 148–149.

Whetstone, G. W. and Schloemer, R. W., 1967, National environmental data-collection systems for water resources development: Internat. Conf. on Water for Peace, Washington, D.C., paper 644, 6 p.

Williams, G. P., 1966, Particle roundness and surface texture effects on fall velocity: Jour. Sed. Petrology, March 1966, v. 36, p. 255–259.

Wischmeier, W. H. and Smith, D. D., 1958, Rainfall energy and its relationship to soil loss: Am. Geophys. Union Trans., v. 39, no. 2, p. 285–291.

——— 1965, Predicting rainfall-erosion losses from cropland east of the Rocky Mountains: U.S. Dept. Agri., Soil Conser. Service Handbook 282.

Wischmeier, W. H., Smith, D. D., and Uhland, R. E., 1958, Evaluation of factors in the soil loss equation: Am. Assoc. of Agr. Engineers, v. 39, no. 8, p. 458–462.

Witzig, B. J., 1943, Sedimentation in reservoirs: Am Soc. Civil Engineers Trans., v. 109, p. 1047.

Wolman, M. G., 1964, Problems posed by sediment derived from construction activities in Maryland: Annapolis, Maryland Water Pollution Control Comm., 125 p.

Yotsukura, N., 1968, The mechanics of dispersion in natural channels: discussion *in* Am. Soc. Civil Engineers Proc., v. 94, HY6, p. 1556–1559.

INDEX

[Italic page numbers indicate major references]

U.S GOVERNMENT PRINTING OFFICE 1970 O—372-206