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Chapter C1

FLUVIAL SEDIMENT CONCEPTS

By Harold P. Guy

Book 3

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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.

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[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;

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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.

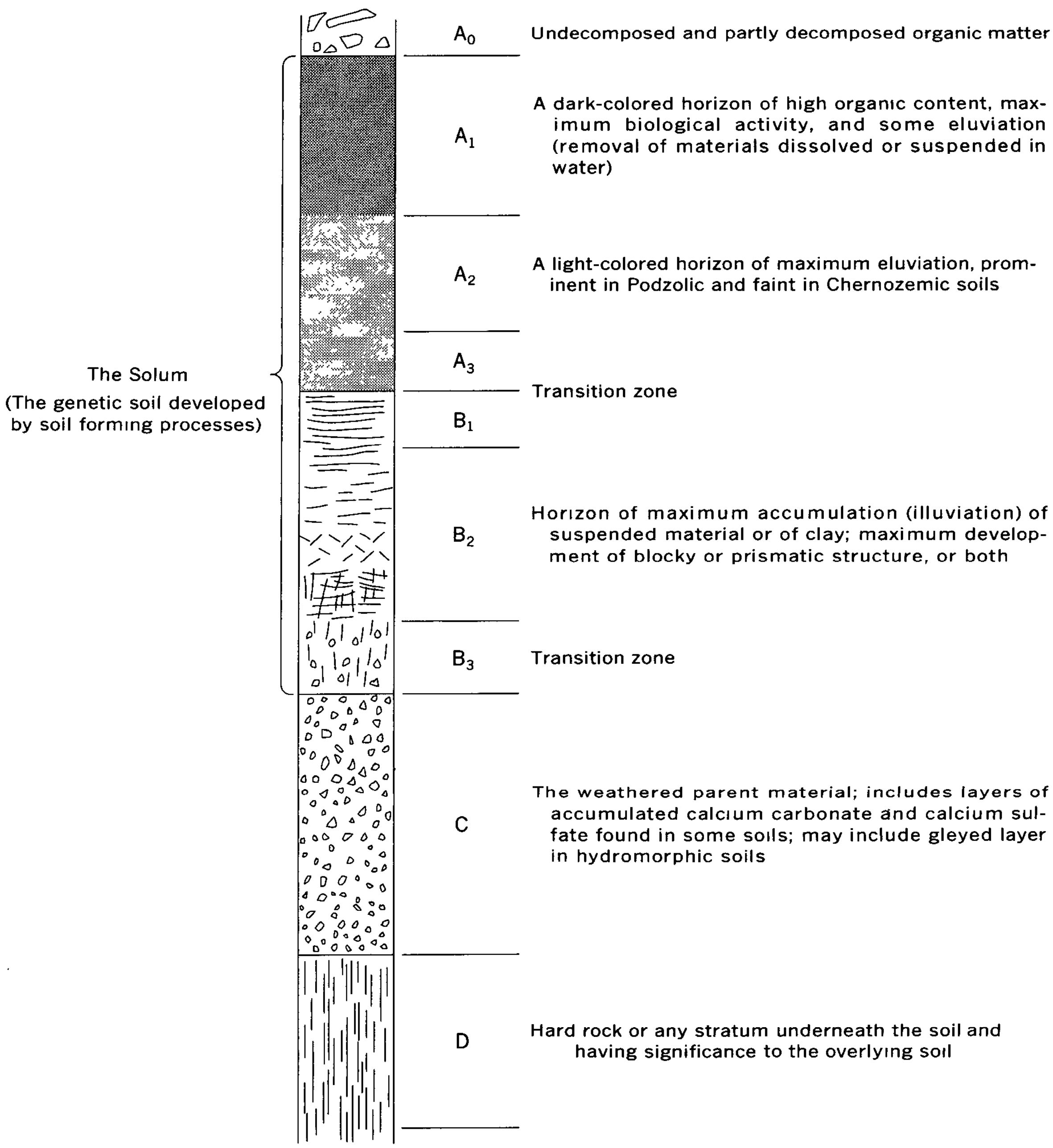


Figure 1.—A hypothetical soil profile of the principal horizons. Every profile has some, but not all, of the indicated features. Modified from Simonson (1957, p. 20).

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

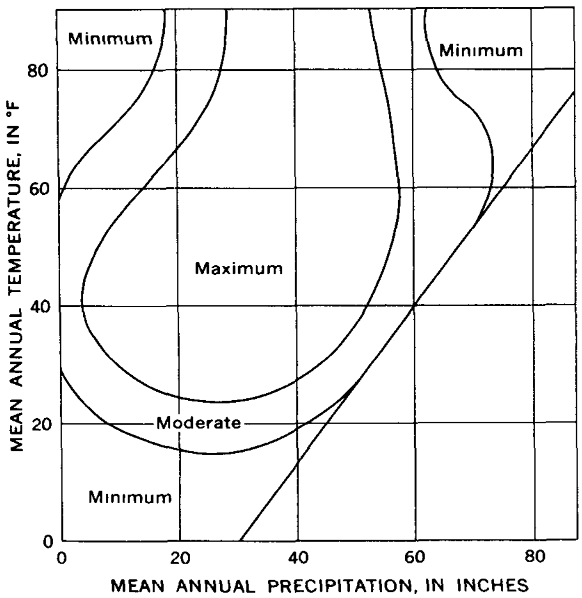


Figure 2.—Relative erosion as related to mean annual temperature and precipitation. Redrawn from Thornbury (1954, p. 60).

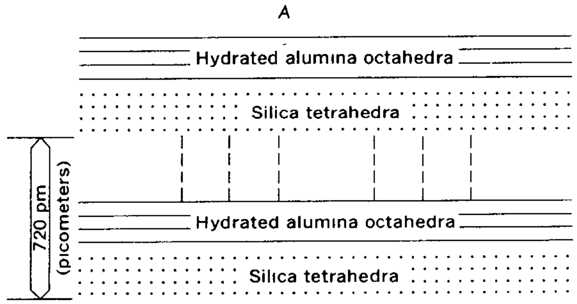
temperature is at or below freezing, where rainfall and temperature are adequate to produce dense vegetative cover, and where rainfall is insufficient at high temperatures to yield runoff because of evapotranspiration. Maximum erosion then occurs at combinations of precipitation and temperature that 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.

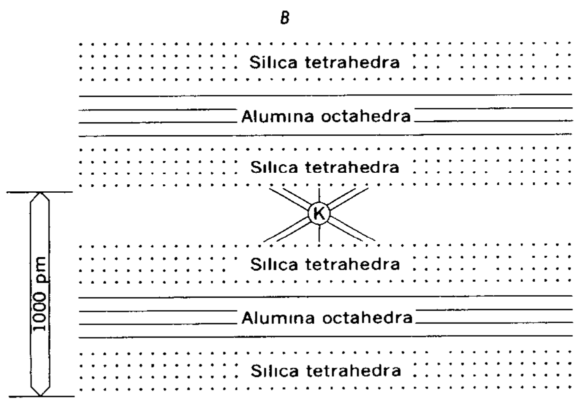
The tractive force required to move a particle against only the interlocking gravitational force can be computed from a hydraulic point of view. The binding forces, on the other hand, are of diverse character and operate by chemical reaction through association of a very large number of very small particles, 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.

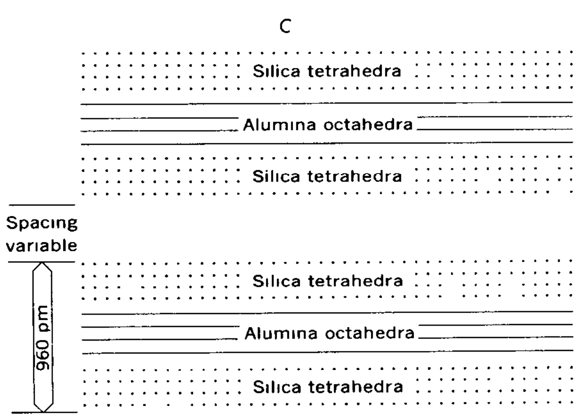
A soil aggregate consists of a grouping of a number of primary particles into a secondary unit. Flocculation occurs when primary particles attract each other upon collision in a water suspension with a low electrokinetic potential. Most such floccules are unstable and break up in other suspensions that 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

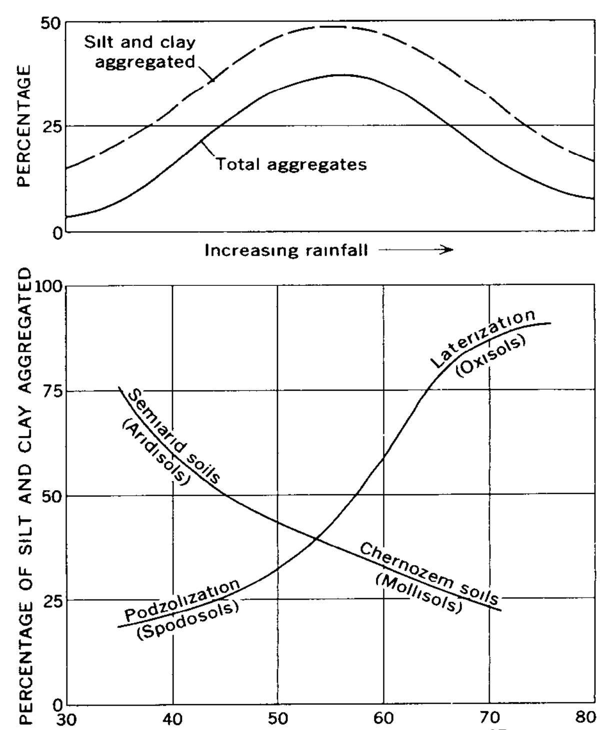


Figure 4.—Relationship of soil aggregation to climatic factors. From Baver (1948, p. 150).

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.

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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,

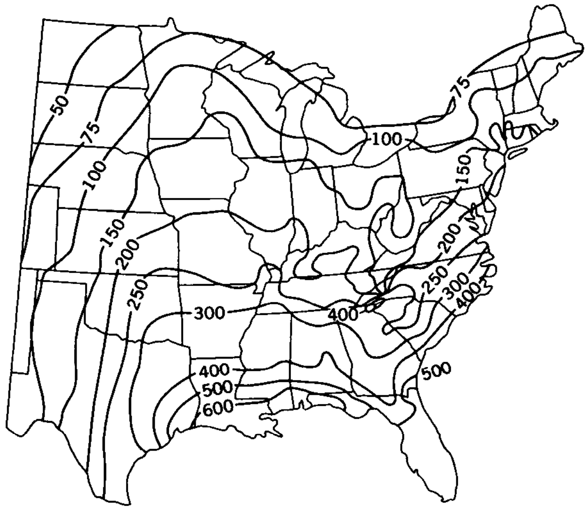

Figure 5.—Mean annual values of Wischmeier's erosion index for the area of the United States east of 105° W.

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-

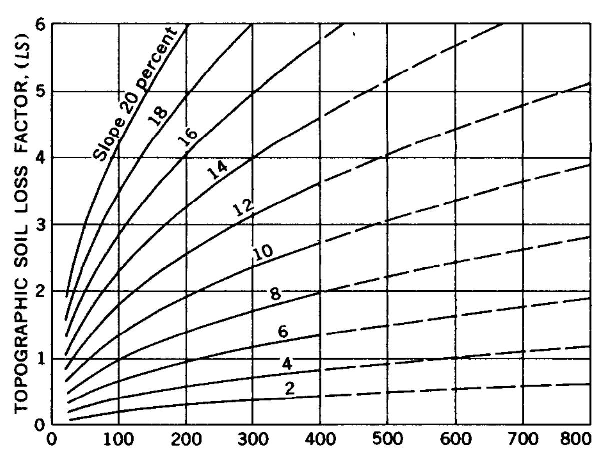

Figure 6.—Relationship of topographic soil-loss factor, LS, to slope length and gradient. The curves indicate that, for a given gradient, soil loss varies with the square root of the slope length.

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 parti-