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SEDIMENT IN THE HYDROLOGIC CYCLE

Rainfall and surface runoff are the agents responsible for the detachment and movement of soil particles on the land surface. These soil particles are referred to as *sediments.* The study of sediment detachment and movement is an important subject in engineering hydrology. Indeed, the subject of sediment transcends engineering hydrology to encompass the related fields of fluvial geomorphology, sediment transport, and sedimentation and river engineering [2, 4, 25, 37].

The study of sediments in the hydrologic cycle can be divided into the following three processes: (1) production, (2) transport, and (3) deposition. These can be linked to the various liquid-transport phases of the hydrologic cycle. At the catchment level, sediment production by soil particle detachment is primarily the result of raindrop impact. Once detachment has taken place, surface runoff acts to transport sediment downslope, first as overland flow (sheet and rill flow), and eventually as stream- and river flow. Deposition of sediment occurs at any point downstream where the kinetic energy of the flow is insufficient to support sediment entrainment in the flowing water.

Sediment production refers to the processes by which sediment is produced, the identification of sediment sources and amounts, and the determination of sediment yields. The source of sediment can usually be traced back to the upland catchments, although these are by no means the only source. In certain cases, streambank erosion in the lower valleys may constitute an important source of sediment.

Sediment from upland catchments is delivered to streams and rivers, wherein sediment transport takes place. Sediment transport refers to the mechanisms by which sediment is moved downstream by flowing water, either in suspension or by rolling and sliding along the river bottom.

The transport of sediment continues in the downstream direction until the flow is no longer able to carry the sediment, at which time sediment deposition occurs. Typically, the first opportunity for sediment deposition is at the entrance to reservoirs and water impoundments, where the flow is decelerated by the action of structures. Deposition is also likely to occur naturally, for instance, downstream of sudden decreases in energy slope or in situations where the capacity of the flow to carry sediment is substantially diminished. In the absence of these natural or human-made features, sediment transport by the flow may continue unabated until it reaches the ocean, at which time the flow loses its kinetic energy and sediment deposition goes on to contribute to delta growth.

This chapter is divided into five sections. Section 15.1 describes sediment properties. Section 15.2 describes sediment production, sediment sources, and sediment yield. Section 15.3 discusses sediment transport, sediment transport formulas, and sediment rating curves, including a brief introduction to sediment routing. Section 15.4 describes sediment deposition in reservoirs. Section 15.5 describes sediment measurement techniques.

15.1 SEDIMENT PROPERTIES

Sediment Formation

Sediments are the products of disintegration and decomposition of rocks. Disintegration includes all processes by which rocks are broken into smaller pieces without substantial chemical change. The disintegration of rocks is caused either by large temperature changes or by alternate cycles of freezing and thawing. Decomposition refers to the breaking down of mineral components of rocks by chemical reaction. Decomposition includes the processes of (1) carbonation, (2) hydration, (3) oxidation, and (4) solution.

Carbon dioxide (CO_2), present in the atmosphere and organic sources, readily unites with water to form carbonic acid (H_2CO_3). Carbonic acid reacts with feldspars to produce clay minerals, silica, calcite, and other relatively soluble carbonates containing potassium, sodium, iron, and magnesium. The addition of water to many of the minerals present in igneous rocks results in the formation of clay minerals such as aluminum silicates. Many secondary minerals are formed from igneous rocks by oxidation, which is accelerated by the presence of moisture in the air. Solution is another important mechanism in the alteration of igneous rock. Oxygen combines with other elements to form sulfates, carbonates, and nitrates, most of which are relatively soluble. The amount (by weight) of dissolved solids carried by streams in the contiguous United States has been estimated at more than 50 percent of the amount of suspended sediment [30].

Particle Characteristics

The characteristics of mineral grains help describe the properties of sediments. Among them are (1) size, (2) shape, (3) specific weight and specific gravity, and (4) fall velocity.

Size. Particle size is a readily measured sediment characteristic. A widely accepted classification of sediments according to size is shown in Table 15-1. Five groups of sizes are included in this table: (1) boulders and cobbles, (2) gravel, (3) sand, (4) silt, and (5) clay. Boulders and cobbles can be measured individually. Gravel-size particles can be measured individually or by sieving. Sand-size particles are readily measured by sieving. A No. 200 screen is used to separate sand particles from finer particles such as silt and clay. Silt and clay particles are separated by measuring the differences in their rate of fall in still water.

Shape. Particle shape is numerically defined in terms of its sphericity and roundness. True sphericity is the ratio of the surface area of a sphere having the same volume as the particle to the surface area of the particle. The practical difficulty of measuring true sphericity has led to an alternate definition of sphericity as the ratio of the diameter of a sphere having the same volume as the particle (i.e., the nominal diameter) to the diameter of a sphere circumscribing the particle. Accordingly, a sphere has a sphericity of 1, whereas all other shapes have a sphericity of less than 1.

TABLE 15-1 CLASSIFICATION OF SEDIMENTS ACCORDING TO SIZE [27]

Class	Size (mm)
Boulders and cobbles	
Very large boulders	4096–2048
Large boulders	2048–1024
Medium boulders	1024–512
Small boulders	512–256
Large cobbles	256–128
Small cobbles	128–64
Gravel	
Very coarse	64–32
Coarse	32–16
Medium	16–8
Fine	8–4
Very fine	4–2
Sand	
Very coarse	2.0–1.0
Coarse	1.0–0.5
Medium	0.50–0.25
Fine	0.250–0.125
Very fine	0.125–0.062
Silt	
Coarse	0.062–0.031
Medium	0.031–0.016
Fine	0.016–0.008
Very fine	0.008–0.004
Clay	
Coarse	0.0040–0.0020
Medium	0.0020–0.0010
Fine	0.0010–0.0005
Very fine	0.0005–0.00025

Roundness is defined as the ratio of the average radius of curvature of the particle edges to the radius of the largest inscribed circle. It refers to the sharpness of the edges of sediment particles and is commonly used as an indicator of particle wear.

In sediment studies, the *shape factor* is often used as an indicator of particle shape:

$$SF = \frac{c}{(ab)^{1/2}} \tag{15-1}$$

in which SF = shape factor and a, b, and c are three orthogonal particle length dimensions. According to Corey [12], a is the longest, b is the intermediate, and c is the shortest length dimension. However, according to McNown and Malaika [32], c is measured in the direction of motion, and a and b are perpendicular to c.

Specific Weight and Specific Gravity. The specific weight of a sediment particle is its weight per unit volume. The specific gravity of a sediment particle is the ratio of its weight to the weight of an equal volume of water. Most sediment particles consist of either quartz or feldspar, which are about 2.65 times heavier than water. Therefore, the specific gravity of sediments is generally considered to be about 2.65. Exceptions are heavy minerals (for instance, magnetite, with specific gravity of 5.18), but these occur rather infrequently.

Fall Velocity. The fall velocity of a sediment particle is its terminal rate of settling in still water. Fall velocity is a function of size, shape, and specific weight of the particle, and the specific weight and viscosity of the surrounding water. For spherical particles, the fall velocity (derived from a balance of submerged weight and drag) can be expressed as follows:

$$w = \left[\frac{4}{3} \frac{g d_s}{C_D} \frac{\gamma_s - \gamma}{\gamma} \right]^{1/2} \tag{15-2}$$

in which w = fall velocity, g = gravitational acceleration, d_s = particle diameter, C_D = drag coefficient (dimensionless), γ_s = specific weight of sediment particles, and γ = specific weight of water.

The drag coefficient is a function of the particle Reynolds number R, defined as:

$$R = \frac{w d_s}{\nu} \tag{15-3}$$

in which ν = kinematic viscosity of the fluid. For particle Reynolds numbers less than 0.1, the drag coefficient is equal to $C_D = 24/R$. Substituting this value of C_D into Eq. 15-2 leads to Stokes' law:

$$w = \left[\frac{g d^2}{18 \nu} \right] \left(\frac{\gamma_s - \gamma}{\gamma} \right) \tag{15-4}$$

For particle Reynolds numbers greater than 0.1, the drag coefficient is still a function of Reynolds number, but the relationship cannot be expressed in analytical form. The relationship of C_D versus R for a wide range of particle Reynolds numbers is shown in Fig. 15-1 [35].

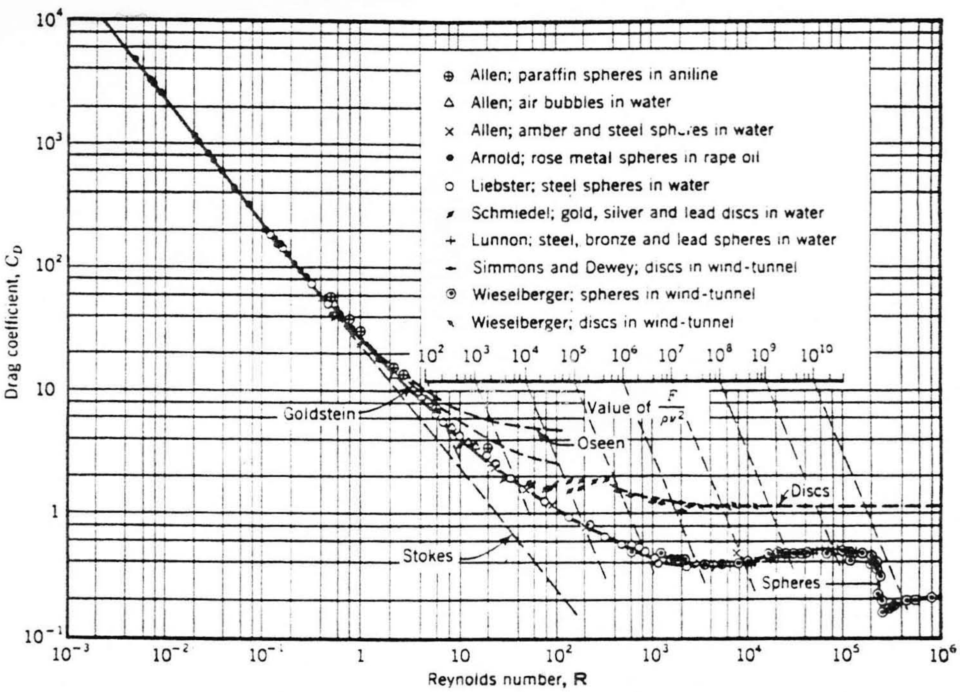

Figure 15-1 Drag coefficient of spheres versus Reynolds number [35].

Since fall velocities vary with fluid temperature and viscosity, two particles of the same size, shape, and specific gravity, falling on two fluids of different viscosity or in the same fluid at different temperatures, will have different fall velocities. To provide a measure of comparison, the concept of *standard fall velocity* was developed [23]. The standard fall velocity of a particle is the average rate of fall that it would attain if falling alone in quiescent water of infinite extent at the temperature of 24°C. Likewise, the standard fall diameter of a particle is the diameter of an equivalent sphere having the same standard fall velocity and specific gravity.

Example 15-1.

Calculate the fall velocity of a spherical quartz particle of diameter $d_s = 0.1$ mm and drag coefficient $C_D = 4$.

Using Eq. 15-2, with $\gamma_s = 2.65$ g/cm^3, and $\gamma = 1.0$ g/cm^3, $g = 9.81$ m/s^2, $d_s = 0.0001$ m: $w = 0.023$ m/s.

Size Distribution of Sediment Deposits

An important property of sediment deposits is the size distribution of its individual particles. Particle size distribution is a key to predicting the behavior of a sediment deposit and estimating its specific weight. A sediment sample containing a wide range of particle sizes is well graded, or poorly sorted. Conversely, a sediment sample consisting of particles in a narrow range of particle sizes is poorly graded, or well sorted.

The size distribution of sediments can be measured in several ways. The coarsest fraction can be separated by direct measurement for boulders and cobbles and by sieving for sands and gravels. For most applications involving sediments in the sand size, the visual accumulation (VA) tube is a fast, economical, and accurate method of determining the size distribution of sediment samples. In the VA tube method, the particles start falling from a common source and become stratified according to their relative settling velocities. At a given instant, the particles coming to rest at the bottom of the tube are of a certain *sedimentation size*, finer than particles that have already settled and coarser than those still remaining in suspension. See [19] for a description of laboratory methods for sediment analysis.

Specific Weight of Sediment Deposits

The specific weight of a sediment deposit is the dry weight of sedimentary material per unit volume. Due to the voids between sediment particles, the specific weight of a sediment deposit is always less than the specific weight of individual particles. A knowledge of the specific weight of a sediment deposit allows the conversion of sediment weights to sediment volumes and vice versa. In particular, the specific weight of a sediment deposit is useful in studies of reservoir storage depletion by deposition of fluvial sediments.

Factors influencing the specific weight of a sediment deposit are (1) its mechanical composition, (2) the environment in which the deposits are formed, and (3) time. Coarse materials, e.g., boulders, gravel, and coarse sand, are deposited with specific weights very nearly equal to their ultimate value and change very little with time. However, fine materials such as silts and clays may have initial specific weights that are only a fraction of their ultimate value.

Lane and Koelzer [28] have developed an empirical relationship to account for the variation of the specific weight of sediment deposits in reservoirs with time. Their relationship is

$$W = W_1 + B \log T \tag{15-5}$$

in which W = specific weight of the deposit after T years; W_1 = initial specific weight of the deposit, measured after 1 y of consolidation; and B = a constant. Table 15-2 shows values of W_1 and B as a function of sediment size and mode of reservoir operation. For mixed deposits, a weighted average of specific weight is appropriate.

Drying or aeration of a sediment deposit helps to accelerate consolidation through removal of the water from the pore spaces. Table 15-3 shows the effect of aeration on the specific weight of sediment deposits for several types of soil mixtures [18].

Example 15-2.

Calculate the specific weight of a sediment deposit in a reservoir after an elapsed time of 50 y, with the sediment always submerged or nearly submerged. Assume the following size distribution: sand, 30%; silt, 45%, clay, 25%.

Using Table 15-2, the specific weights for the various sizes are: sand, 93 lb/ft^3; silt, 75 lb/ft^3; clay, 57 lb/ft^3. Therefore, the weighted average is: $W = (93 \times 0.30) + (75 \times 0.45) + (57 \times 0.25) = 75.9$ lb/ft^3.

TABLE 15-2 CONSTANTS FOR ESTIMATING SPECIFIC WEIGHT OF RESERVOIR SEDIMENT DEPOSITS, EQ. 15-5 (lb/ft³) [28]

Mode of Reservoir Operation	Sand		Silt		Clay	
	W_1	B	W_1	B	W_1	B
Sediment always submerged or nearly submerged	93	0	65	5.7	30	16.0
Normally a moderate reservoir drawdown	93	0	74	2.7	46	10.7
Normally considerable reservoir drawdown	93	0	79	1.0	60	6.0
Reservoir normally empty	93	0	82	0.0	78	0.0

Note: 1 lb/ft^3 = 157.1 N/m^3.

TABLE 15-3 RANGE IN SPECIFIC WEIGHT OF SEDIMENT DEPOSITS (lbs/ft³) [18]

Soil Description	Permanently Submerged	Aerated
Clay	40–60	60–80
Silt	55–75	75–85
Clay-silt mixture	40–65	65–85
Sand-silt mixture	75–95	95–110
Clay-silt-sand mixture	50–80	80–100
Sand	85–100	85–100
Gravel	85–125	85–125
Poorly-sorted sand and gravel	95–130	95–130

Note: 1 lb/ft^3 = 157.1 N/m^3.

15.2 SEDIMENT PRODUCTION

The presence of sediment in streams and rivers has its origin in soil erosion. Erosion encompasses a series of complex and interrelated natural processes that have the effect of loosening and moving away soil and rock materials under the action of water, wind, and other geologic factors. In the long term, the effect of erosion is the denudation of the land surface, i.e., the removal of soil and rock particles from exposed surfaces, their transport to lower elevations, and eventual deposition.

The rate of landscape denudation can be quantified from a geological perspective. For instance, the number of centimeters of denudation per 1000 y can be used as a measure of the erosive activity of a region. Geologic measures of landscape denudation appear insignificant when compared to the typical timespan of human activity, say 25 to 100 y. However, the quantities of sediment moved may be important when considering the impact that sediment loads have on the operation and design life of reservoirs and hydraulic structures.

At the outset of the study of sediment production, a distinction should be made between the amount of sediment eroded at the source(s) and the amount of sediment delivered to a downstream point. *Gross sediment production* refers to the amount of sediment eroded and removed from the source(s). *Sediment yield* refers to the actual delivery of eroded soil particles to a given downstream point. Since eroded particles may be deposited before they reach the downstream point of interest, sediment yield quantities are generally less than gross sediment production quantities. The ratio of sediment yield to gross sediment production is the *sediment-delivery ratio (SDR)*.

Gross sediment production is commonly measured in terms of weight of sediment per unit drainage area per unit time—for instance, metric tons per hectare per year, or tons per acre per year. Sediment yield is expressed in terms of weight per unit time past a certain point—for instance, metric tons per day at the catchment outlet.

Normal and Accelerated Erosion

According to the timespan involved, erosion can be classified as (1) normal, or geologic, and (2) accelerated, or human-induced. Normal erosion has been occurring at variable rates since the first solid materials formed on the surface of the earth. Normal erosion is extremely slow in most places and is largely a function of climate, parent rocks, precipitation, topography, and vegetative cover. Accelerated erosion occurs at a much faster rate than normal, usually through reduction of vegetative cover.

no ¶

Deforestation, overgrazing, overcultivation, forest fires, and the systematic destruction of natural vegetation result in accelerated erosion.

Sediment Sources

According to its source, erosion can be classified as (1) sheet erosion, (2) rill erosion, (3) gully erosion, and (4) channel erosion. *Sheet erosion* is the wearing away of a thin layer on the land surface, primarily by overland flow. *Rill erosion* is the removal of soil by small concentrations of flowing water (rills). *Gully erosion* is the removal of soil from incipient channels that are large enough so that they cannot be removed by normal cultivation. *Channel erosion* refers to erosion occurring in stream channels in the form of streambank erosion or streambed degradation. For practical purposes, a distinction is made between upland and channel erosion. *Upland erosion* is mostly made up of sheet and rill erosion, whereas channel erosion encompasses all other sediment sources, specifically excluding sheet and rill erosion.

Upland Erosion and the Universal Soil Loss Equation

In the United States, the prediction of upland erosion amounts (i.e., sheet and rill erosion) is commonly made by the universal soil loss equation (USLE), developed by the USDA Agricultural Research Service in cooperation with USDA Soil Conservation Service and certain state experiment stations.

The universal soil loss equation is [43]:

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The universal soil loss equation is [43]:

$$A = RKLSCP \tag{15-6}$$

in which A = (annual) soil loss due to sheet and rill erosion in tons per acre per year; R = rainfall factor; K = soil erodibility factor; L = slope-length factor; S = slope-gradient factor; C = crop-management factor; and P = erosion-control-practice factor.

Rainfall Factor. When factors other than rainfall are held constant, soil losses from cultivated fields are shown to be directly proportional to the product of the storm's total kinetic energy E and its maximum 30-minute intensity I. The product EI reflects the combined potential of raindrop impact and runoff turbulence to transport dislodged soil particles.

The sum of EI products for a given year is an index of the erosivity of all rainfall for that year. The rainfall factor R is the average value of the series of annual sums of EI products. Values of R applicable to the contiguous United States are shown in Fig. 15-2.

Soil Erodibility Factor. The soil erodibility factor K is a measure of the resistance of a soil surface to erosion. It is defined as the amount of soil loss (in tons per acre per year) per unit of rainfall factor R from a *unit plot.* A unit plot is 72.6 ft long, with a uniform lengthwise gradient of 9 percent, in continuous fallow, tilled up and down the slope.

Values of K for 23 major soils on which erosion plot studies were conducted since 1930 are listed in Table 15-4. Soil erodibility factors for other soils have been estimated by comparing their characteristics with those of the 23 soils listed in Table 15-4. A method for determining the soil erodibility factor based on soil characteristics has been proposed by Wischmeier et al. [44].

Slope-length and Slope-gradient Factors. The rate of soil erosion by flowing water is a function of slope length (L) and gradient (S). For practical purposes, these two topographic characteristics are combined into a single topographic factor (LS). The topographic factor is defined as the ratio of soil loss from a slope of given length and gradient to the soil loss from the unit plot (of 72.6 ft length and 9 percent gradient). Figure 15-3 shows values of LS as a function of slope length and gradient.

Crop-management Factor. The crop-management factor C is defined as the ratio of soil loss from a certain combination of vegetative cover and management practice to the soil loss resulting from tilled, continuous fallow. Values of C range from as little as 0.0001 for undisturbed forest land to a maximum of 1.0 for disturbed areas with no vegetation. Values of C for cropland are estimated on a local basis. Table 15-5 shows values of C for permanent pasture, grazed forest land, range, and idle land. Table 15-6 shows values of C for undisturbed forest land.

Erosion-control-practice Factor. The erosion-control-practice factor P is defined as the ratio of soil loss under a certain erosion-control practice to the soil loss resulting from straight-row farming. Practices for which P have been established are contouring and contour strip-cropping. In contour strip-cropping, strips of sod or meadow are alternated with strips of row crops or small grains. Values of P used for contour strip-cropping are also used for contour-irrigated furrows. Table 15-7 shows values of P for contour-farmed terrace fields.