United States
Department of
Agriculture

Soil
Conservation
Service

National Engineering Handbook

Section 3

Sedimentation

Chapter 2

Sediment Properties

Contents

	Page
General	2-1
Derivation of sediments	2-2
Disintegration	2-2
Decomposition	2-2
Carbonation	2-3
Hydration	2-3
Oxidation	2-3
Solution	2-3
Soil formation	2-3
Particle characteristics	2-5
Size	2-5
Shape	2-5
Specific gravity	2-5
Size distribution of sediments	2-8
Fine-grain separation	2-8
Sieve separation	2-8
Fall velocity	2-8
Bulk properties	2-11
Sorting	2-11
Glacial and other ice-action deposits	2-11
Alluvial fans	2-11
Alluvial deposits	2-11
Beach deposits	2-11
Colluvial deposits	2-11
Marine sediments	2-11
Eolian deposits	2-12
Lacustrine deposits	2-12
Chemical deposits and evaporites	2-12
Texture	2-12
Volume-weight	2-13
Stability of sediment deposits	2-13
Porosity	2-13
Permeability	2-14
References	2-17

Figures

Page

2–1. Dimensions required for sphericity and roundness calculations 2-7
2–2. Grain-size distribution graph 2-9
2–3. Settling rates of sediment particles 2-10

Tables

Page

2–1. Chemical weathering products of common rock-forming silicate minerals 2-4
2–2. Common minerals: their hardness, specific gravity, and frequency of occurrence in average igneous rocks and sediments 2-6
2–3. Sediment grade scales 2-7
2–4. Volume-weight of reservoir sediment 2-14
2–5. Approximate average porosity of various formations 2-14
2–6. Permeability and porosity as related to grain-size distribution 2-15

Chapter 2
Sediment Properties

General

Sediments are the products of disintegration and decomposition of rocks. Material becomes detached and is transported to a deposition site where it may be affected by solution, cementation, consolidation, or biological action.

The physical properties of sediments depend on a number of factors, including composition, texture, and structure of the original formation; topography; type of weathering; and sorting (Lobeck 1939, pp. 63-80).[1] The greatest variety of minerals and textures in sediment comes from the weathering of igneous rocks, especially from this disintegration in semiarid and arid climates. These conditions have produced great volumes of sediment containing much coarse material, including boulders, especially along mountain fronts and in intermontane valleys. These deposits commonly contain a relatively high proportion of unaltered minerals such as feldspars, amphiboles, pyroxenes, and micas. Sediments produced by erosion in more humid and deeply weathered areas generally have a finer texture and a higher proportion of minerals produced by chemical weathering.

Small grains of certain minerals resistant to chemical weathering, such as zircon, quartz, rutile, tourmaline, topaz, and ilmenite, remain in sediment relatively unchanged. These detrital mineral suites may reveal the source rock type (Krumbein and Sloss 1963, p. 108). Feldspars, the most common minerals in igneous rock (Pettijohn 1957, p. 122), are much less stable and less common in sediments. In humid climates feldspars are relatively easily decomposed to form products including clay minerals, silica, and oxides of aluminum.

[1]*See the list of references at the end of each chapter for more information on these sources.*

Weathering and deterioration of rocks are considered the primary mechanisms of sediment formation. The processes and agents causing rock deterioration are many and diverse, and only a brief summary is presented in this handbook as a guide to the proper interpretation of sediments. Participants in sedimentation surveys should understand that rock formations—igneous, sedimentary, and metamorphic, either consolidated or unconsolidated—are subject to major deterioration and alteration at or near the earth's surface.

Weathering of rocks is an adjustment to a new environment. Intrusive igneous and metamorphic rocks are exposed to weathering when erosion removes the formations covering them. The forces of weathering attack volcanic rocks and sedimentary formations exposed at the earth's surface. The rate of rock deterioration depends on many factors, including composition and structure of the formation, climate, topography (especially slope), nature of vegetal cover, and elevation. The processes of weathering have been studied by many investigators, and much information is available.[2]

Disintegration

Disintegration (physical disruption) includes all processes by which rocks are broken into smaller pieces without much chemical change. Rocks are broken either into pieces containing all their original minerals in a relatively unaltered state or into grains, each consisting of an original mineral. The result may be splitting of blocks from a formation or disintegration to sand or gravel.

Large and rapid temperature changes can disrupt rock masses. For example, forest fires can heat exposed rock rapidly, thus fracturing and fragmenting it.

Frost is a major agent in the disintegration of rocks. Water collects in voids and openings in rocks and, on freezing, increases about 9 percent in volume. This transformation of water from a liquid to a solid state can dislodge fragments of rock as large as 10 feet in maximum dimension, as found along cliffs bordering Devil's Lake, Wis. (Leet and Judson 1958, p. 81). Angular fragments a few inches across are a common result of frost action on rocks.

Relief of pressure is a disruptive force where weathering and subsequent transportation remove a load from underlying rock formations, especially on steep slopes. The same effect can be produced by landslides that remove an overlying load.

Diastrophism of any type disrupts rocks. It can cause new joint systems, widen preexisting joints, or produce movement along a fracture during an earthquake. Folding of rock formations over long periods of time can also be a disruptive force.

Products of mechanical disintegration range from large boulders to sand. Basic igneous rocks commonly yield sand and gravel composed of calcic feldspars and relatively unaltered ferromagnesian minerals.

Moving water and ice are powerful disruptive forces on rock formations in several environments. These forces include wave action along shores of seas and lakes, abrasion of the banks and beds of streams, and scouring and plucking by glacial ice. The atmosphere is also a disruptive force, especially in arid regions where rocks in exposed positions are subject to attack by winds carrying abrasive mineral particles.

Biological agents have some disruptive effects on rocks, including widening of crevices by root growth, pitting of rock surfaces by lichens, and burrowing by some animals.

Decomposition

All rocks located at or near the surface of the earth are subject to decomposition as well as disintegration. Decomposition is the breaking down of mineral components of rocks by chemical reaction. Most decomposition occurs above the groundwater table, but the processes of weathering extend down hundreds of feet in desert regions and in some regions of high rainfall.

Igneous rocks are generally susceptible to chemical attack, since they are definitely out of equilibrium with the environment near the earth's surface. Twenhofel has found that, on the average, 100g of igneous rock acquires through decomposition 5.3g of carbon dioxide, 2g of water, 0.7g of carbon, and about 1g of oxygen (Mason 1956, p. 130). The resultant rocks have lower specific gravity and higher porosity than the unweathered igneous rocks. Averages of many analyses by Clarke (1924)

[2] *Authorities on weathering processes and products include Lobeck (1939), Reiche (1950), Pettijohn (1957), and Leopold, Wolman, and Miller (1964).*

and others indicate that the weathering of igneous rocks has produced sedimentary rocks in about the following proportions: shale, 82 percent; sandstone, 12 percent; and limestone, 6 percent.

Carbonation

Carbon dioxide (CO_2) is one of the most important and most common weathering agents. It comes from the atmosphere and from organic sources. It readily unites with water to form the weak acid H_2CO_3 (carbonic acid). Carbonic acid reacts with feldspars to produce clay minerals, silica, calcite, and other relatively soluble carbonates containing potassium, sodium, iron, and magnesium. The common carbonate rocks are limestone, dolomite, and marls.

Hydration

The addition of water to many of the minerals of igneous rocks results in the formation of clay minerals, which are hydrous aluminum silicates. Many minerals formed by hydration become dull earthy masses that contrast with their former hard, crystalline nature. Hydration also nearly doubles the volume of material (Lobeck 1939, p. 76). The transformation of feldspar to kaolinite is an example.

Oxidation

Through oxidation, many secondary minerals are formed from igneous rocks. The oxides of aluminum and iron are among the most stable. The oxidation of rocks in air is accelerated in the presence of moisture. Ferrous silicates in pyroxenes, amphiboles, and olivine are oxidized by air and water to hematite (ferric oxide, Fe_2O_3). The oxidation of iron is marked by color changes from green or black to red, yellow, or brown. Oxygen combines with other elements to form sulfates, carbonates, and nitrates, most of which are relatively soluble.

Solution

Solution is important in the alteration of igneous rock. Some minerals, such as quartz and the accessory minerals, are relatively insoluble. An accumulation of quartz grains thus becomes sand or sandstone. Clays and shales contain decomposition products of the feldspars and other less common primary silicates. Some of the silica from any of the silicates may be removed in solution (see table 2–1). The ground water and streams contain more silica in solution in areas of igneous rock than in sedimentary terranes. This is so partly because quartz, which is more common in sediments, is less soluble than the other common silicates and partly because less stable silicates are somewhat desilicated in the earlier cycle of sedimentation. The basic igneous rocks, such as basalt and gabbro, contain much silica even if they lack free quartz. The silica in solution and the colloidal-size silica are carried away and may be redeposited in crevices as veins of quartz or may become a cementing material filling interstices or even a replacement mineral as in silicified wood. It has been estimated that the weight of dissolved solids carried by streams in the conterminous United States is more than 50 percent of the weight of the suspended sediment carried (Leifeste 1974).

The carbonates are important solution products. Some carbonates reach the ocean and become important constituents of marine deposits as a result of the chemical or biochemical action that produces limestones and marls.

Soil Formation

Most sediment with which SCS is concerned results from erosion of soil that has taken many centuries to form. Weathered rock and soil differ in that soil contains organic as well as mineral matter and has more than one layer (horizon) roughly parallel to the land surface. Soil formation begins when material weathered from bedrock develops two or more distinguishable horizons. Most soil profiles include three principal horizons, identified by the letters A, B, and C (Simonson 1957).

Table 2–1.—Chemical weathering products of common rock-forming silicate minerals[1]

Mineral	Composition	Important decomposition products	
		Minerals	Others
Quartz	SiO_2	Quartz grains	Some silica in solution.
Orthoclase	$K(AlSi_3O_8)$	Clay Quartz (finely divided)	Some silica in solution. Potassium carbonate (soluble).
Albite (sodic plagioclase)	$Na(AlSi_3O_8)$	Clay Quartz (finely divided)	Some silica in solution. Sodium and calcium carbonates (soluble).
Anorthite (calcic plagioclase)	$Ca(Al_2Si_2O_8)$	Calcite (from Ca)	
Biotite Augite Hornblende	Fe, Mg, Ca silicates of Al	Clay Calcite Limonite Hematite Quartz (finely divided)	Some silica in solution. Carbonates of calcium and magnesium (soluble).
Olivine	$(Fe, Mg)_2SiO_4$	Limonite Hematite Quartz (finely divided)	Some silica in solution. Carbonates of iron and magnesium (soluble).

[1]From Leet and Judson (1958). Reprinted by permission of Prentice-Hall, Inc., Englewood Cliffs, N.J.

Various characteristics of mineral grains composing sediments have different effects on the formation and subsequent development of deposits. Size, shape, hardness, specific gravity, chemical composition, and degree of weathering of the mineral grains affect the rate and place of deposition and the nature of the deposits ultimately formed. Table 2–2 lists some of the common minerals and their hardness, specific gravity, and relative abundance.

Size

Size is an important particle characteristic that is readily measured. Bulk properties tend to vary with particle size in a roughly predictable manner. In fact, size alone has been found to describe sediment deposits adequately for many practical purposes.

Various organizations have adopted different size classifications to meet their particular needs. Four are shown in table 1–1 of Chapter 1 in Section 8, Engineering Geology, SCS National Engineering Handbook. A grade scale based on Wentworth's classification (Wentworth 1922) was recommended by the American Geophysical Union (1947) and is reproduced in table 2–3.

Five groups of sizes are presented in this table: boulders and cobbles, gravel, sand, silt, and clay. The largest size is uncommon but is easily measured. Gravel-size particles are more important than boulders and cobbles and are transported in some streams as bedload. Gravel can be measured directly by diameter or volume or by sieving. Sand-size sediment is common and is easily sized by sieving. The finest screen, No. 200, can be used for accurate size separation of sand and silt. Silt and the other fines, the clays, are best separated by measuring their rate of fall in a fluid. Silt and clay together make up most of the suspended load in streams, and they are usually distributed uniformly throughout the depth of the stream. Clay-size particles are important in their effect on density currents and on the change in volume-weight of sediment deposits during consolidation.

Shape

The various shapes of sediment particles are formed in numerous ways. Some shapes, such as the roundness of river and beach pebbles or the facets of wind-abraded particles, indicate the environment in which they formed. Other shapes express mineralogic characteristics; examples are the curving shards of volcanic glass and the unworn crystals of many resistant minerals.

Shape is defined numerically by sphericity and roundness (fig. 2–1). 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. Sphericity is also expressed as d_n/d_s, where d_n is the nominal diameter (diameter of a sphere having the same volume as the particle) and d_s is the diameter of a circumscribing sphere. A sphere has a sphericity of 1, and all other shapes have a sphericity of less than 1 (Pettijohn 1957, p. 56).

Roundness describes the sharpness of the edges and corners of a particle and is an indication of the wear the particle has received. Roundness is defined as the average radius of curvature of the edges, r_a, divided by the radius of the maximum inscribed circle, R.

Specific Gravity

The specific gravity of a mineral is the ratio of its weight to the weight of an equal volume of water. Most sediment consists of quartz or feldspar particles, which are about 2.65 times heavier than water, so a specific gravity of 2.65 is generally considered characteristic of sediment. Heavy minerals (for example, magnetite with specific gravity of 5.18), of course, are found in many sediments, but they make up such a small percentage that their importance is minor. For SCS geologists, the chief value of heavy minerals in sediment deposits is that they provide a means of identifying the sediment source.

Table 2-2.—Common minerals: their hardness, specific gravity, and frequency of occurrence in average igneous rocks and sediments.

Mineral	Mohs scale of hardness	Specific gravity	Frequency of occurrence	
			Average igneous rock[1]	Sediment[2]
			Percent	*Percent*
Feldspars	6–6	2.6–2.8	59.5	15.6
Hornblende and pyroxene	5–6	2.9–3.3	16.8	—
Quartz	7	2.65	12.0	34.8
Micas	2–4	2.7–3.1	3.8	15.1
Titanium minerals	5–6	3.4–5.5	1.5	trace
Clay minerals		2.0–3.0	—	14.5
Dolomite	3.5–4	2.8–2.9	—	9.1
Calcite	3	2.7	—	4.2
Limonite	1–5.5	3.4–4.0	—	4.0
Apatite	4.5–5	3.2	0.6	0.4
Gypsum	1.5–2	2.2–2.4	—	1.0
Other			5.8	1.3
Total			100.0	100.0

[1]Clarke (1924).

[2]Leith and Mead (1915).

Table 2–3.—Sediment grade scales[1]

Diameter			
In millimeters	In micrometers	In inches	Class
4,096–2,048		160–80	Very large boulders
2,048–1,024		80–40	Large boulders
1,024–512		40–20	Medium boulders
512–256		20–10	Small boulders
256–128		10–5	Large cobbles
128–64		5–2.5	Small cobbles
64–32		2.5–1.3	Very coarse gravel
32–16		1.3–0.6	Coarse gravel
16–8		0.6–0.3	Medium gravel
8–4		0.3–0.16	Fine gravel
4–2		0.16–0.08	Very fine gravel
2.0–1.0	2,000–1,000		Very coarse sand
1.0–0.5	1,000–500		Coarse sand
0.50–0.25	500–250		Medium sand
0.25–0.125	250–125		Fine sand
0.125–0.062	125–62		Very fine sand
0.062–0.031	62–31		Coarse silt
0.031–0.016	31–16		Medium silt
0.016–0.008	16–8		Fine silt
0.008–0.004	8–4		Very fine silt
0.004–0.0020	4–2		Coarse clay
0.0020–0.0010	2–1		Medium clay
0.0010–0.0005	1–0.5		Fine clay
0.0005–0.00024	0.5–0.24		Very fine clay

[1]From American Geophysical Union (1947).

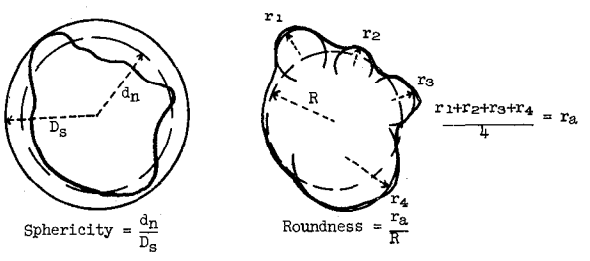


Figure 2–1.—Dimensions required for sphericity and roundness calculations.

One of the most important properties of sediment deposits is the particle-size distribution of the mineral grains. The distribution is important in predicting the behavior of sediment and estimating its specific weight. A number of precautions must be taken in studying deposits in the field and selecting samples for laboratory analysis. Laboratory studies cannot supply answers to many field problems. Problems such as selecting the beds or deposits to be sampled and determining the origin of deposits and the rate of deposition must be solved in the field. Field and laboratory data—the nature of the sediment and its texture, as well as its relationship to other formations, to soils, and to land use—must be interpreted.

The size frequency distribution of a sediment can be measured in a number of ways. The coarsest fraction is differentiated by direct measurement of gravels or larger sizes and by sieving sands. Fine-grained sediments can be separated by elutriation (the determination of settling velocity in a sediment-liquid mixture) or by microscopic examination. Detailed methods of analyzing sediments are presented by Guy (1969).

Fine-Grain Separation

One method of fine-grain separation is by timing the settling rate of sediment particles in a column of water. A suspension of the sediment sample is treated with a deflocculant such as sodium carbonate, is thoroughly mixed, and is then put into a graduated cylinder containing a column of water 800 mm high. After 10 min the upper part of the suspension is drawn off with a siphon. The coarse sediment containing grains 1/16 mm and larger remains at the bottom. This process is usually repeated about four times to achieve a clean separation. The coarse and fine separates can then be treated and studied separately.

A popular modification of the elutriation technique involves use of a bottom withdrawal tube (Howard 1948). The apparatus consists of a graduated glass cylinder with a constriction and a valve at the bottom through which the coarse particles are withdrawn. From the separation thus obtained, a cumulative curve showing size distribution can be plotted on Form SCS-ENG-353, grain-size-distribution graph (fig. 2–2). Other modifications include use of hydrometers to measure the density of the suspension at various time intervals and pipettes to withdraw fine fractions at definite time intervals.

Sieve Separation

Coarse grains (larger than 0.062 mm) are ordinarily separated by sieves having mesh openings corresponding to the grain sizes measured. The U.S. standard sieve series is based on a 200-mesh screen with a diameter of 0.074 mm. Sets of sieves with openings larger than this diameter include 0.125-mm, 0.25-mm, 0.50-mm, 2-mm, and 4-mm sizes (Twenhofel and Tyler 1941, p. 46). Grains of various sizes can be separated by this method according to the scales shown in table 2-3. The dry sample is put in the top sieve of a stack and shaken. Usually 10 min in a mechanical shaker is enough for good size separation. The material caught on each screen is weighed, and the results are expressed as a percentage of the total sample weight. For uniformity in classification of sediment texture, SCS personnel should use table 2–3 as a standard. Table 2–3 shows texture classes as standardized by the American Geophysical Union (1947).

Fall Velocity

The settling rate of particles is influenced primarily by the size, shape, and specific gravity of the particles and by the viscosity and temperature of the medium. Of these characteristics, grain size is the most important for a given fluid. The settling rates of various minerals and aggregates vary widely.

Figure 2–3 has been developed from calculations of settling velocities and laboratory measurements to illustrate the fall velocity of particles in still water at 25° C. The viscosity of water varies with temperature; settling rates decrease as temperature falls and increase as temperature rises.

Form SCS 357
9-64

LABORATORY NO. _____

MATERIALS TESTING REPORT	U. S. DEPARTMENT of AGRICULTURE SOIL CONSERVATION SERVICE	SOIL CLASSIFICATION

PROJECT and STATE: Clear Creek, Site #2, Any State			SAMPLE LOCATION: Borrow Area A
FIELD SAMPLE NO.: 151.1	DEPTH: 3 - 5'	GEOLOGIC ORIGIN: Alluvium (Flood Plain)	
TYPE OF SAMPLE: Disturbed	TESTED AT: SML - Lincoln	APPROVED BY:	DATE: 1/21/78
SYMBOL: GW-GC	DESCRIPTION: Well graded clayey gravel		

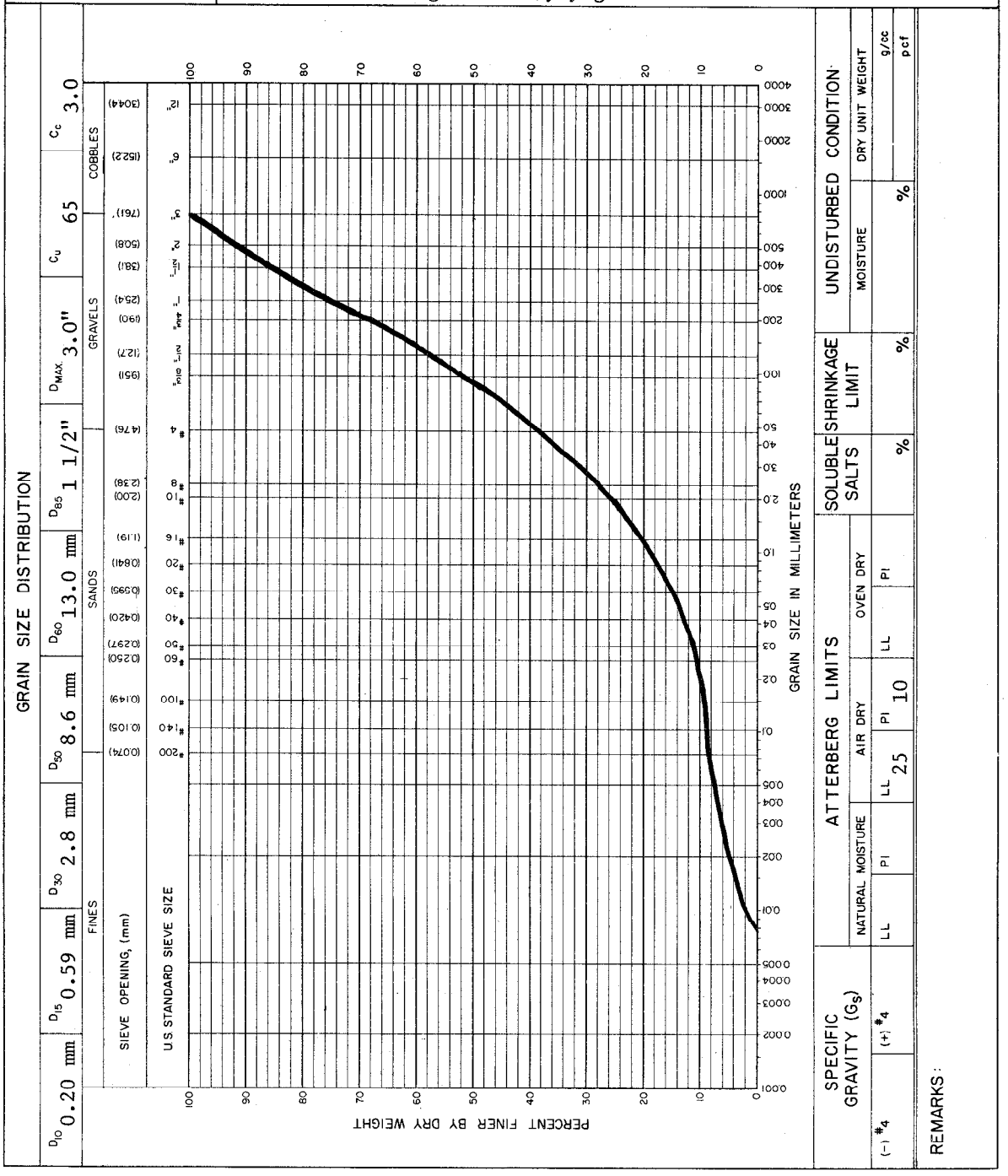
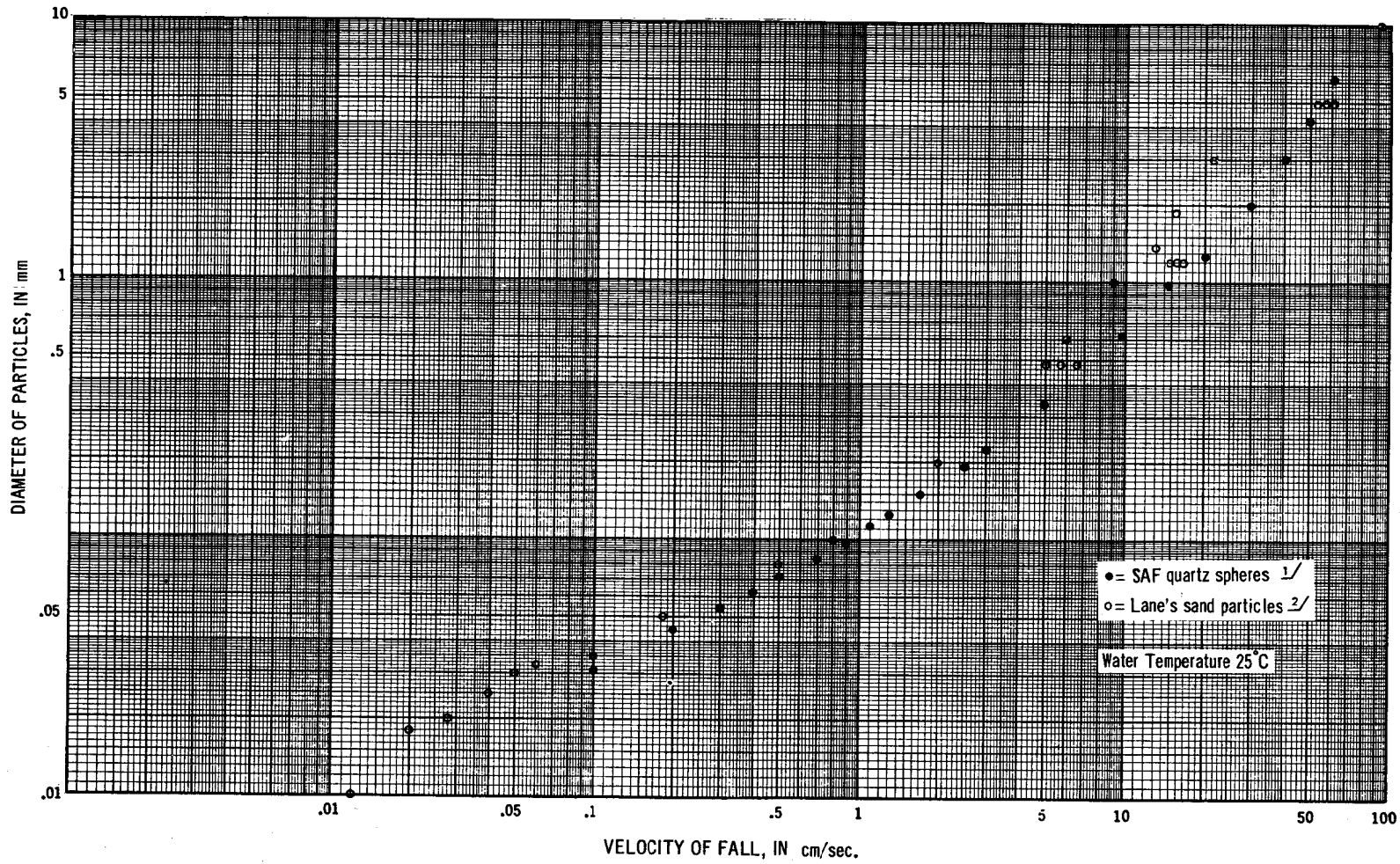


Figure 2–2.—Grain-size distribution graph.



1/ INTER-AGENCY SEDIMENTATION PROJECT AT ST. ANTHONY FALLS LABORATORY (1957)

2/ LANE (1938)

Figure 2-3.—Settling rates of sediment particles.

This discussion applies primarily to unconsolidated sediments, although sedimentary deposits range from loose deposits of mineral grains to consolidated formations of similar composition that have been lithified into indurated rocks such as sandstone, limestone, and shale. Sources of sediments studied in SCS watershed investigations can be rocks, weathered zones, or soils. SCS studies of eroded and transported debris primarily involve unconsolidated, mostly uncemented sediments. The characteristics of all such formations depend on the nature and arrangement of the individual grains in the aggregate. Hence, sorting, environment of deposition, mineral species, water-holding and -transmitting capacity, and thickness of deposits affect the characteristics of the sediment deposit.

Sorting

The degree of sorting in a sediment deposit is determined by the similarity or dissimilarity of the component particles. Similarity can apply to a number of characteristics, including size, shape, specific gravity, and mineral or chemical composition. In most sediment studies the classification refers to size distribution. The engineering term "well graded" means poorly sorted and that the deposit contains a number of size grades.

The following classification includes a list of environments in which sediment deposits form, arranged approximately from the most poorly sorted to the best sorted deposits. This classification, like many others, is subject to many exceptions, some of which are indicated in the following summary descriptions. Sharp distinctions between the various groups of sediment deposits cannot be made; they all grade into the adjacent groups. Authorities on sediment sorting include Grabau (1913), Krumbein and Sloss (1963), Pettijohn (1957), and Wentworth (1922).

Glacial and Other Ice-Action Deposits

Deposits formed by glacial action are among the most poorly sorted of all sediment deposits. Glacial till, left by melting glaciers, contains fragments of all sizes, from large boulders to finely ground fragments called rock flour. Moraines and glacial outwash deposits may be more uniform, but they almost always contain much gravel, as well as sand, silt, and clay.

Alluvial Fans

A wide range of sizes is characteristic of piedmont or alluvial-fan deposits; hence, they form one of the groups of poorly sorted sediments. A lower gradient at the foot of steep slopes causes rapid deposition of most of the load of vigorous and rapid streams. Large rock blocks and boulders are commonly mixed with pebbles, sand, silt, and clay with little or no stratification.

Beach Deposits

Sorting of beach (littoral) deposits is usually poor. These deposits are primarily along shorelines and harbors along seacoasts, but they are also along the shorelines of large lakes. Locally, the sediments may be relatively well sorted and uniform in areas where conditions are stable, but in general the alternate rising and falling tides and the alternate dominance of tidal and river currents cause deposition of poorly sorted sediments. Some authorities on littoral deposits are Caldwell (1950), Grabau (1913), Mason (1950), and Trask (1950).

Alluvial Deposits

Sediments composing alluvial deposits vary greatly in size and other characteristics. Alluvial deposits can range in area from a narrow strip in a small stream valley to a great plain such as the High Plains deposit that extends east from the Rocky Mountains. In the upstream reaches of a valley where stream action is vigorous, alluvial deposits generally are coarse and poorly sorted. In the middle reaches of most streams, the coarsest and most poorly sorted parts of the alluvial deposit form in the channel. This coarse deposit is distributed to some extent over the valley bottom as the stream meanders. The deposits that occur farther downstream usually are better sorted and contain a relatively high percentage of fine sediments—fine sands, silts, and clays.

Colluvial Deposits

These products of upland erosion consist of heterogeneous materials of any particle size that accumulate on the lower part or base of slopes. Colluvium is transported there by gravity (talus), sheetwash, soil creep, and mudflows.

Marine Sediments

Marine sediments range from heterogeneous

gravel and boulder deposits to vast widespread oozes in the deep sea. Deposits having the narrowest range of size distribution form where conditions of deposition remain uniform or vary only slightly for long periods. These conditions persist along slowly advancing or retreating shorelines, where great deposits of uniform sands are formed and widespread deposits of clay are laid in the deeper water. Where conditions are favorable for chemical precipitation, with or without the action of biological agents, thick and extensive deposits of carbonates accumulate. These deposits eventually become crystalline limestone or similar rock.

Eolian Deposits

Most sedimentary deposits of eolian origin are among the better sorted groups of terrestrial deposits. The following four groups of elolian deposits are recognized:

Loess.—Loess is one of the best sorted and most texturally uniform of the terrestrial deposits. Loess deposits can range from a featheredge to hundreds of feet thick and consist mainly of silt-size particles that have been transported by wind. Topographic irregularities such as a line of bluffs in a valley increase deposition. Loess deposits cover wide areas in the United States. Since they are mostly unconsolidated, they are subject to rapid erosion and gully development and they contribute to accelerated sediment deposition in reservoirs and stream channels and on flood plains.

Dunes.—Sand dunes are windblown deposits of grains moved mostly by traction or saltation, especially in semiarid and arid areas. Dunes form in areas such as lake shores, seacoasts, glacial plains, and lake beds as well as in deserts. They are generally well sorted and consist predominantly of fine- to medium-grain sands. As a result of the transporting power of the wind, the silt- and clay-size particles may be carried long distances, leaving the sand to accumulate as slow-moving dunes.

Desert pavement or wind-lag deposits.—These deposits form in many desert areas where wind removes the sand and finer textured material. The resulting surface is a thin residual concentration of wind-polished, closely packed pebbles, gravel, and other rock fragments.

Volcanic dust.—Wind caries great quantities of volcanic dust long distances after volcanic explosions. This material is well sorted; the particles that travel the farthest are all silt and clay size.

Lacustrine Deposits

Sediment deposition in lakes and reservoirs produces some of the best sorted nonmarine sedimentary deposits. The bulk of the sediment in most lakes—that found in all the larger and deeper parts of the basins, where currents are not vigorous—is almost entirely silt and clay size. These deposits are, therefore, well sorted and fine grained. The coarser and generally more poorly sorted lacustrine sediments are common along shore zones, where wave action is vigorous and coarse detritus is available, and in upstream segments, where inflowing streams deposit their coarse material.

Chemical Deposits and Evaporites

Sediment deposited from solution and evaporation is the best sorted of all sedimentary deposits. These deposits may consist of mineral crystals of almost uniform size. If organisms are incorporated in the deposit, the shells or skeletons add pieces of different sizes, reducing the degree of sorting.

Texture

The size, shape, and arrangement of the particles composing a sediment deposit determine its texture. Differences in the texture of the many types of sediment deposits cause relatively large variations in the damage that results from accelerated deposition. Coarse sediments of alluvial fans consist chiefly of gravel and boulders and cause major damage if deposited on agricultural land. Overbank flood deposits produce damage that usually increases as the texture of the deposited sediment becomes coarser. Deposits of clays and silts usually have some fertility, but they may bury crops or impede drainage if thick enough. Regardless of their texture, sediment deposits occupy valuable space in reservoirs, obstruct bridge and culvert openings, decrease stream channel and ditch capacity, and cause many other types of damage.

Differences in the texture of sediment deposits control or modify the uses for which the deposits are suited in agriculture, industry, and construction. Sand and gravel formations are the most important as aquifers and are essential materials in concrete. The uses of sand in glass manufacture, of clays in the ceramics industry, and of combinatic

of sediment deposits having a variety of textures in construction are beyond the scope of this chapter, but some properties affecting earth-dam construction are described in Section 8, Engineering Geology, SCS National Engineering Handbook.

Cumulative curves (fig. 2–2) are used for presenting data on particle-size distribution. Histograms relating pyramidal curves to texture are sometimes used. Because the distribution of grain size in most samples is not symmetrical, the amount of skewness is also shown by the pyramids. This is well explained by Pettijohn (1957).

Volume-Weight

One of the most important properties of sediment deposits is weight per unit volume, or volume-weight. Volume-weight, as it applies to measurement of eroded sediment, sediment in transport, and sediment deposits in place, has primary importance in the sedimentary cycle.

Information on the unit weight of sediment deposits for construction or other purposes reflects many variations in properties. For example, a cubic foot of quartz, which has a specific gravity of 2.65, weighs about 165 lb. Similarly, a cubic foot of solid magnetite, specific gravity 5.2, weighs 324 lb. Most sediment deposits, in contrast, weigh about 25 to 125 lb/ft^3 when water free (USDA 1978). The volume-weight of sediment deposits is largely determined by the proportion of voids present. If the sediment is below the water surface, the voids are filled chiefly with water. If the sediment is exposed to the atmosphere, there are fewer voids and they are filled chiefly with air or a combination of air and water, depending on rainfall, seepage, and other factors.

Volume-weight has been measured in conjunction with many types of investigations, including construction projects, geological surveys, sedimentation surveys of reservoirs and valleys, and soil surveys.

Table 2–4 shows the average volume-weight of some typical reservoir sediment deposits (USDA 1978). This table is arranged in two parts: Part A shows the weight per cubic foot, determined by laboratory analysis, of sediment samples from reservoirs in which the water level is near the spillway crest most of the time, and Part B gives the same information for undisturbed samples from reservoirs in which sediment is exposed to the air during repeated low water levels.

Stability of Sediment Deposits

A high degree of angularity of individual sediment grains of silt size or larger promotes stability. A loose aggregation of angular grains is more stable in steeper slopes than an aggregation of more rounded grains. Similarly, angular particles in earthfills increase resistance to slumping and shear. Aggregates of mostly silt- and clay-size particles usually have predominantly angular or platy pieces, but their stability in a fill is determined more by water content and overburden pressure than by the shape of the grains. Deposits of loess, which may be tens or hundreds of feet thick and are composed of highly angular silt-size particles, tend to stand in nearly vertical faces. Deposits of more rounded grains, such as alluvial or coarse-grained eolian deposits, have lower angles of repose and are usually less stable. Deposits of platy pieces, which have an abundance of grains with two long and one short dimension, are also readily susceptible to sliding.

Porosity

Porosity has been described by Graton and Fraser (1935), Meinzer (1923), Todd (1959), and Tolman (1937) in connection with the volume and movement of ground water. According to Graton and Fraser, an arrangement of spheres providing about 49 percent pore space has the greatest porosity. This arrangement, however, is unstable. The most stable arrangement of spheres of uniform diameter provides about 25 percent pore space, but it is not found in natural sediments. Porosity exceeding 50 percent has been measured in natural sedimentary deposits.

Meinzer (1923) defines porosity in a rock or soil as the property of containing interstices or voids. The percentage of pore space is determined by the distribution of fine grains between coarser grains, the shape of the particles, and their arrangement. Grains of silt and clay size occupying spaces between sand and gravel particles can reduce porosity significantly. Both porosity and stability of sediment deposits are affected by the shape of their mineral grains. Many studies have shown that fine-grained sediments are subject to far more compaction and decrease in volume than are deposits of sand or larger grains.

Table 2–4.—Volume-weight of reservoir sediments

Name	Physiographic section	Average volume-weight (lb/ft³)	Main stream	Nearest city
		A. Submerged sediment		
Lake Throckmorton	Redbeds Plains	31.6	Brazos River	Throckmorton, Tex.
Lake Ft. Phantom Hill	Red Hills	36.0	Clear Fork, Brazos	Abilene, Tex.
Grand Saline Reservoir	Forested Coastal Plain	38.7	Saline River	Grand Saline, Tex.
Lake Woodland	Coastal Plain	39.9	Camp Creek	DeSoto County, Miss.
Madison Lake	Central Till Plain	49.9	Ohio River	Madison, Ohio
Crab Orchard Lake	Central Till Plain	47.5	Big Muddy River	Carbondale, Ill.
White Rock	Black Prairie	49.0	Trinity River	Dallas, Tex.
Lake Williams	Appalachian Mountains	49.1	Susquehanna River	York, Pa.
Moran Reservoir	Osage Plains	49.2	Osage River	Moran, Kans.
Lake Issaqueena	Piedmont Plateau	50.9	Savannah River	Clemson, S.C.
Mountain Lake	Ozark Mountains	54.8	St. Francis River	Patterson, Mo.
Herman Lage Pond	Central Till Plain	54.8	Missouri River	Aspinwall, Iowa
Lake Harris	Piedmont Plateau	59.9	Black Warrior River	Tuscaloosa, Ala.
		B. Sediment frequently aerated		
Franklinton Reservoir	Piedmont Plateau	67.0	Sallie River	Franklinton, N.C.
White Manganese No. 6	Appalachian Mountains	87.5	Tallapoosa River	Cartersville, Ga.
Medina Lake	Edwards Plateau	75.0	San Antonio River	San Antonio, Tex.
Backbone Lake	Driftless Area	75.1	Mississippi River	Stawberrt Pt., Iowa
Lee Johnson Pond	Coastal Plain	76.1	Tallahatchie River	Holly Springs, Miss.
Castlewood Reservoir	High Plains	77.5	Arkansas River	Denver, Colo.
West Frankfort Reservoir	Central Till Plain	78.2	Big Muddy River	West Frankfort, Ill.
Gerber Lake	Sacramento Valley	78.3	Sacramento River	Corning, Calif.
Loring Lake	Coastal Plain	85.1	Mississippi River	Zwolle, La.
Cobb Creek No. 3	Osage Plains	90.0	Washita River	Colony, Okla.
Lake Medicine	Redbeds Plains	98.5	Chickaskia River	Medicine Lodge, Kans.

Table 2–5 illustrates the range in average porosity of various materials. This table does not take into account the degree of cementation or the fact that although a fine-grained deposit such as a clay may have high porosity, it permits little movement of water.

Table 2–5.—Approximate average porosity of various formations[1]

Formation	Porosity (percent)
Clay	45
Silt	40
Sand	35
Gravel	25
Shale, sandstone	18
Limestone	10
Granite, basalt	1

[1]From Leopold, Wolman, and Miller (1964).

Permeability

The permeability of sediments varies widely. Permeability is extremely low in clay materials, even though they may have high porosity and be water saturated. The interstices between the clay particles are small enough for molecular attraction to hold water tightly. Permeability is highest for coarse, clean gravel. Table 2–6 indicates the relation of permeability and porosity to grain-size distribution.

Table 2–6.—Permeability and porosity related to grain-size distribution[1]

Percent smaller than indicated grain size by weight							Porosity (percent)	Coefficient of permeability (gal/day/ft²)
2.0 mm	1.0 mm	0.5 mm	0.25 mm	0.125 mm	0.062 mm	0.005 mm		
—	99.4	98.4	95.4	89.7	65.0	21.0	58.2	0.0002
—	—	—	97.5	95.6	94.6	49.3	55.5	0.2
—	93.9	67.8	18.4	3.2	1.3	—	46.6	60.0
84.3	69.1	48.9	29.4	13.0	6.0	1.5	26.3	150.
77.6	59.9	34.1	10.6	0.9	[2]0.2	—	28.9	1,000.
77.7	58.1	38.4	19.3	9.6	5.3	0.9	25.0	2,095.
24.3	15.7	6.3	1.1	0.4	[2]0.2	—	23.4	4,200.
20.2	19.1	16.9	12.0	7.5	4.8	1.2	25.1	20,663.
8.4	0.5	0.4	0.3	0.2	[2]0.1	—	38.0	90,000.

[1]After Wenzel (1942).

[2]Includes clay (<0.005 mm).

References

American Geophysical Union. 1947. Report of the subcommittee on sediment terminology. Am. Geophys. Union Trans. 28:936–938.

Brown, Carl B. 1950. Sediment transportation. *In* Engineering hydraulics, p. 769–857. Hunter Rouse, ed. John Wiley & Sons, Inc., New York.

Caldwell, Joseph M. 1950. Sedimentation in harbors. *In* Applied sedimentation, p. 290–299. Parker D. Trask, ed. John Wiley & Sons, Inc., New York.

Clarke, F. W. 1924. The data of geochemistry. U.S. Geol. Surv. Bull. 770, 841 p.

Grabau, A. W. 1913. Principles of stratigraphy. A. G. Seiler & Co., New York, 1,185 p.

Graton, L. C., and H. J. Fraser. 1935. Systematic packing of spheres—with particular relation to porosity and permeability. J. Geol. 43(8):785–909.

Guy, H. P. 1969. Laboratory theory and methods for sediment analysis. *In* U.S. Geol. Survey techniques of water-resources invest., Book 5, Chap. C1, 58 p.

Happ, Stafford C., Gordon Rittenhouse, and G. C. Dobson. 1940. Some principles of accelerated stream and valley sedimentation. U.S. Dep. Agric. Tech. Bull. 695, 134 p.

Howard, C. S. 1948. Laboratory experiences with the bottom withdrawal tube method of size analysis. Fed. Inter-Agency Sediment. Conf. Proc., Denver, Colo., p. 235–243.

Kraus, E. H., W. F. Hunt, and L. S. Ramsdell. 1959. Mineralogy. McGraw-Hill Book Co., New York, 686 p.

Krumbein, W. C., and F. J. Pettijohn. 1938. Manual of sedimentary petrography. D. Appleton-Century Co., New York, 549 p.

Krumbein, W. C., and L. L. Sloss. 1963. Stratigraphy and sedimentation. 2d ed. W. H. Freeman & Co., San Francisco, Calif., 660 p.

Lane, E. W. 1938. Notes on the formation of sand. Am. Geophys. Union Trans. 19(pt. 1): 505–508.

Leet, L. D., and S. Judson. 1958. Physical geology. 2d ed. Prentice-Hall, Inc., Englewood Cliffs, N.J., 502 p.

Leifeste, D. K. 1974. Dissolved-solids discharge to the oceans from the conterminous United States. U.S. Geol. Surv. Circ. 685, 8 p.

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

Lobeck, A. K. 1939. Geomorphology. McGraw-Hill Book Co., New York, 731 p.

Mason, B. 1956. Principles of geochemistry. John Wiley & Sons, Inc., New York, 276 p.

Mason, Martin A. 1950. Geology in shore-control problems. *In* Applied sedimentation, p. 276–290. Parker D. Trask, ed. John Wiley & Sons, Inc., New York.

Meinzer, O. E. 1923. Outline of ground-water hydrology. U.S. Geol. Surv. Water-Supply Pap. 494, 71 p.

Pettijohn, F. J. 1957. Sedimentary rocks. 2nd ed. Harper & Bros., New York, 718 p.

Reiche, Parry. 1950. A survey of weathering processes and products. N. Mex. Univ., Publ. Geol. 3, 95 p.

Simonson, Roy W. 1957. What soils are. *In* Soil, U.S. Dep. Agric. Yearb. Agric., p. 17–31. U.S. Govt. Printing Off., Washington, D.C.

Subcommittee on Sedimentation, Interagency Committee on Water Resources. 1957. The development and calibration of the visual accumulation tube. Rep. 11, St. Anthony Falls Hydraulic Laboratory, Minneapolis, Minn., table 11, p. 99.

Todd, D. K. 1959. Ground water hydrology. John Wiley & Sons, Inc., New York, 336 p.

Tolman, C. F. 1937. Ground water. McGraw-Hill Book Co., New York, 593 p.

Trask, Parker D. 1950. Dynamics of sedimentation. *In* Applied sedimentation, p. 3–40. John Wiley & Sons, Inc., New York.

Twenhofel, W. H. 1950. Principles of sedimentation. 2d ed. McGraw-Hill Book Co., New York, 673 p.

Twenhofel, W. H., and S. A. Tyler. 1941. Methods of study of sediments. McGraw-Hill Book Co., New York, 183 p.

U.S. Department of Agriculture, Agricultural Research Service. 1978. Sediment deposition in U.S. reservoirs, summary of data reported through 1975. In cooperation with Sediment. Comm., Water Resour. Counc. Misc. Publ. 1362, 82 p.

Wentworth, C. K. 1922. A method of measuring and plotting the shapes of pebbles. U.S. Geol. Surv. Bull. 730–C, p. 91–114.

Wenzel, L. K. 1942. Methods for determining permeability of water-bearing materials. U.S. Geol. Surv. Water-Supply Pap. 887, 192 p.