

1 Physical Properties of Primary Particles

Joseph M. Skopp

1.1 INTRODUCTION

This chapter discusses the following physical properties of the primary particles: particle density, particle shape, particle size distribution and surface area. In addition, two soil properties related to packing are also presented: bulk density and porosity. The definitions and ideas behind these properties have built into them the concept or assumption of discrete primary particles as the primary soil constituent. If organic matter or amorphous materials and cementing agents are abundant, then the importance of primary particles is reduced.

1.2 PARTICLE DENSITY (ρ_p)

1.2.1 DEFINITION

The particle density represents one of the fundamental soil physical properties. Particle density is defined as the mass of soil particles divided by the volume occupied by the solids (i.e., excluding voids and water). Typical values for soils range from 2.5–2.8 Mg/m³, with 2.65 Mg/m³ being representative of many soils. Particle density provides few insights into soil physical processes. Consequently, one frequently overlooks the errors associated with particle density. However, its value enters into calculations of more useful soil properties, such as porosity and particle size distribution.

1.2.2 TYPICAL VALUES — MINERAL, WHOLE SOIL

Each individual soil mineral has a characteristic particle density. Values for different minerals can be found in Klein and Hurlbut (1985). Quartz, a predominant soil mineral, has a value of about 2.65 Mg/m³ which is why this value is frequently given as representing all soils. In contrast, gypsum has a value of 2.32 Mg/m³, biotite a value of 2.80–3.20 Mg/m³ and hematite a value of 4.80–5.30 Mg/m³. The particle density of a soil is an average for the distribution of soil minerals present.

A determination of particle density may be made for individual minerals, size fractions or the soil as a whole. Organic matter removal takes place in order to reduce variation. With standard procedures and removal of organic matter, a propagated error of less than $\pm .01$ Mg/m³ is possible. (Propagated error is the combination of all instrument errors when making a determination of a soil physical property.) Generally, determinate errors (i.e., biases) play a greater role in the analysis of particle density than indeterminate errors (i.e., random errors).

Perhaps the most interesting question in the determination of soil particle density is the role of organic matter (with a typical density of 1.0 Mg/m³) in surface horizons. Most standard methods remove organic matter; thus the particle density reflects only the mineral phase. This value is the best value for use in particle size analysis, but may not be the best value for calculating porosity. Including organic matter means that changes in soil management may change this particle density.

1.2.3 METHODS — WATER PYCNOMETER, AIR PYCNOMETER, OTHER

Three methods of determining particle density will be examined. The most common determination of soil particle density uses a pycnometer or some variation. A pycnometer is any device which can be made to retain a reproducible or measurable volume. The soil sample is introduced into the pycnometer and the displaced volume of a fluid is determined. Water is typically the displaced fluid, but air can also be used.

The water pycnometer method generally requires the removal of organic matter prior to use. This avoids problems with trapped air and increased variability. The water pycnometer method requires good temperature control.

An alternative air pycnometer procedure uses a gas as the displacing fluid and the ideal gas law to calculate the volume of solids. The principles of an air pycnometer are straightforward, but care must be taken to prevent temperature changes. The air pycnometer does not require the removal of organic matter, which is particularly valuable when the total porosity (or total void space) or air-filled porosity is required without knowing soil bulk density or water content.

A less common method of determining particle density uses a vibrating tube which is filled with a solution or suspension. The resonant frequency of the vibrating tube provides a very precise means for determining the density of the suspension.

EXAMPLE 1.1 Particle density of whole soil. Is the particle density obtained from standard techniques suitable for determining total porosity?

The standard equation to calculate porosity requires the particle density. Standard methods require the removal of organic matter to carry out this procedure. The effective particle density including organic matter can easily be 10% less, which can result in a significant error.

1.3 PARTICLE SHAPE

1.3.1 DEFINITIONS — SHAPE FACTORS, FRACTALS

Particle shape influences specific surface, as well as particle size, analysis. It is also expected that particle shape has a strong influence on the packing of particles and soil strength, as well as transport properties. Unfortunately, particle shape is difficult to measure and few determinations exist in the literature.

A variety of terms exist to define particle shape. Some of the definitions require reference to a figure of an ideal particle (Figure 1.1). Let L , B and T represent the length, breadth and thickness of a particle. Then Heywood (1947) defines the following:

$$\text{Flatness Ratio} = B/T$$

$$\text{Elongation Ratio} = L/B$$

Other dimensionless terms are:

$$\text{Sphericity} = \frac{\text{surface area of equivalent sphere}}{\text{actual surface area}}$$

$$\text{Circularity} = \frac{\text{circumference of circle with the same area}}{\text{actual perimeter}}$$

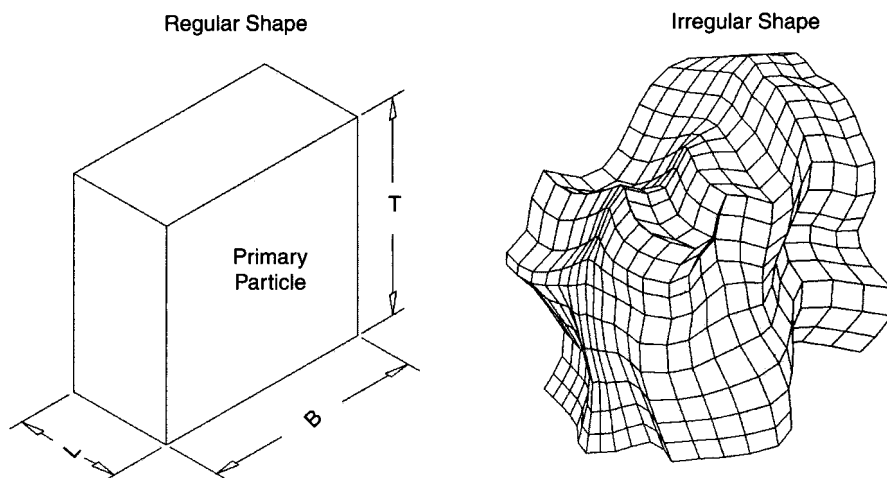


FIGURE 1.1 Ideal individual soil particle.

TABLE 1.1
Shape factors for three-dimensional physical models
 with $L = d$, where d is a diameter or edge length, and
 $h =$ disc height or rod length.

Geometry	Area	C_2	Volume	C_3
Sphere	πd^2	π	$\pi d^3/6$	$\pi/6$
Disc or cylinder	$(\pi dh + \pi d^2/2)$	$(\pi h/d + \pi/2)$	$\pi d^2 h/4$	$\pi h/4d$
Cube	$6d^2$	6	d^3	1
Square rod	$2d^2 + 4dh$	$2 + (4h/d)$	$d^2 h$	h/d

$$\text{Rugosity} = \frac{\text{actual perimeter}}{\text{circumference of circumscribing circle}}$$

These last two terms work best for a two-dimensional image or the projection of the particle outline onto a flat surface.

A microscopic picture of soil describes either the solids or the voids. A simple picture of the solids might be a sphere or cube, while that of the voids might be a cylindrical tube or a slit between two flat surfaces. These pictures, or physical models, of the soil make it possible to deduce relations describing surface area, packing, water retention or water movement.

It is possible to apply these ideas to models that are not simple geometrical figures. Here, a characteristic length replaces the edge length of a cube and a dimensionless shape factor (C_k) describes the deviation of the shape from simple geometries (Table 1.1). These two factors are used to describe the area and volume as:

$$A = C_2 L^2 \quad \text{and} \quad V = C_3 L^3 \quad [1.1]$$

where L is the characteristic length and C_2 or C_3 is the shape factor for area or volume, respectively. For a cube, if L equals d (the length of an edge), find: $C_2 = 6$ and $C_3 = 1$. Examples of C_2 and C_3 for other shapes are in Table 1.1. Two particles differing in size may or may not have the same shape. Typically, weathering or size reduction changes not only the total dimension but also the shape factor.

1.3.2 METHODS

At least three methods exist to examine particle shape. First is direct observations under a microscope or using an image analysis system. Commercial image analysis systems exist that automatically provide shape factors or similar properties.

The second method is an indirect technique from the variation of viscosity with the concentration of suspended particles. Increasing the solids concentration results in deviations from a pure liquid viscosity. These deviations are dependent on particle geometry as well as concentration of the suspension. The viscosity of the suspension (η) and the viscosity of pure solvent (η_s) usually behave as follows:

$$\eta/\eta_s = 1 + Kf \quad [1.2]$$

where f is the volume fraction of suspended material and K is an empirical constant, which is a shape parameter (Einstein, 1906). For spheres, $K = 2.5$ is a constant and changes with the shape of the particle. Kahn (1959) applied this technique to examine the shape of clay particles. Similar techniques (Egashira and Matsumoto, 1981) provide estimates of a/b (the ratio of major to minor axis) for montmorillonite (200–300), kaolinite (15–25) and mica (10–20).

The third method of particle shape analysis uses the scattering characteristics of light passing through a soil suspension and relies on measurements of the angles of scattered light. Instruments commercially available for particle size analysis can be used.

EXAMPLE 1.2 Shape factor for perimeter. Define the shape factor (C_1) for perimeter (P) of a particle. Show two ways to estimate C_1 from the examination of microscopic images.

By analogy with Equation [1.1], $P = C_1L$ where L is the characteristic length or diameter. The shape factor is:

$$C_1 = P/L$$

The perimeter may be estimated directly. Or the area (A) of the particle's image is used to get the perimeter, $P = A/L$, and then:

$$C_1 = A/L^2$$

L must be carefully defined (e.g., minimum diameter, maximum diameter or some average). A can be a projected area on the image, or the three-dimensional surface of a particle. Note that calculations using these two equations will not necessarily yield the same shape factor.

1.4 PARTICLE SIZE DISTRIBUTION (PSD)

1.4.1 DEFINITIONS (INCLUDING CONCEPT OF TEXTURE)

Particle size distribution (PSD) is the most fundamental physical property of a soil and defines the soil texture. The particle sizes present and their relative abundance sharply influence most physical properties.

Soil particle size (or effective diameters) provide the basis for a classification system. A range of diameters may be given a special designation (e.g., 2.0 mm to 1.0 mm is very coarse sand). Typically the ranges form a logarithmic scale with particles in a given size range termed soil separates. The size boundaries vary with country or discipline. Comparisons of the names given to a size range are given elsewhere (Gee and Bauder, 1979; Sheldrik and Wang, 1993).

The phrase, equivalent diameters, is used to emphasize the role of the measurement technique in determining particle size. *If identical particles are measured by different techniques, they may appear to have different diameters. It is conceivable that two soils with identical PSDs (as determined by a single method) will show differences in other physical properties resulting from distinct particle shapes.*

Defining the diameter of an irregularly shaped particle is not a trivial task. A single parameter, the diameter, characterizes a smooth sphere. The symmetry of the sphere and its smoothness mean that no other information need describe it. As soon as a distortion of the sphere occurs (i.e., into a jelly bean) then at least three diameters are possible. Some particle size analysis methods may orient the particle into a preferred direction (e.g., settling of a particle in a liquid). Other methods (e.g., image analysis) may observe several possible orientations.

1.4.2 TYPICAL DISTRIBUTIONS

Typical data for a variety of soils are presented in Table 1.2. Interpretation of particle size analysis data requires either the drawing of graphs or the calculation of summary coefficients, which are discussed in Section 1.4.7.

Graphs of a PSD typically select the dependent variable as either the cumulative fraction up to a size or incremental fraction of soil in a size interval. The incremental fraction (F_i) is usually the mass of particles within a size interval (X_{i-1} to X_i) divided by the total mass of solids with the index i specifying the size interval. The cumulative fraction (G_i) is the sum of all fractions for particle sizes less than the X_i value.

A typical graphical expression of PSD uses the logarithm of particle diameter (Figures 1.2 and 1.3). The shape and position of the graph provides qualitative information about the soil texture. Soils frequently show a log normal distribution of particle sizes so that a graph of fraction versus the logarithm of particle diameters appears to be normally distributed.

1.4.3 DISPERSION AND FRACTIONATION

Particle size analysis (or mechanical analysis) consists of isolating various particle sizes or size increments and then measuring the abundance of each size. Most methods accomplish this in two steps. First, the soil is dispersed, or separated into individual primary particles. Second, the dispersed sample is fractionated, or the amounts of each size interval are measured.

There are three objectives of dispersion: 1) removal of cementing agents, 2) rehydration of clays, and 3) the physical separation of individual soil particles. It is sufficient to recognize that organic matter and amorphous minerals are the primary cementing agents. When either of these are present in large amounts (e.g., Histosols or Oxisols), then dispersion may be difficult or meaningless. Generally, soil dispersion occurs using a combination of chemical and mechanical means.

It is important to recognize that the fraction of soil that is a single size cannot be determined. What is detected is the fraction of soil within a particular particle size interval or the cumulative fraction of all particles less than a given size. The use of sieves typically determines the mass fraction (mass of particles in a size interval divided by total mass). Microscopic counting results in a number fraction (number of particles in a size interval divided by total number of particles). Photometric techniques typically determine the area fraction. Other methods result in volume or line fractions, depending on the sensing procedure. Thus, while all the methods are capable of observing PSDs, not all the methods provide results that are equal or directly comparable.

TABLE 1.2
Particle size distribution of soil samples representing a variety of soil types from the United States. Mass percent of total sample for the indicated size class.

Soil*	Sand					Silt				Clay	
	Very Coarse 2-1 mm	Coarse 1-0.5 mm	Medium 0.5-0.25 mm	Fine 0.25-0.10 mm	Very Fine 0.1-0.05 mm	Coarse 50-20 μ	Medium 20-10 μ	Fine		2-0 μ	
								10-5 μ	5-2 μ		
Anthony	18.05	13.71	17.68	12.93	8.92	7.41	2.69	2.20	1.37	15.04	
Ava	0.53	0.56	1.25	0.82	0.67	12.80	21.69	15.71	9.63	30.63	
Chalmers	0.74	0.62	1.67	1.38	2.52	19.42	20.18	11.18	7.44	34.89	
Davidson	0.71	2.38	6.52	6.02	3.39	3.32	4.83	4.08	7.43	61.32	
Fanno	8.45	4.87	2.40	9.96	9.06	5.92	4.27	4.05	5.56	46.46	
Kalkaska	0.19	1.79	47.99	36.26	5.19	1.34	1.01	1.33	0.18	4.67	
Mohave	15.25	11.30	12.40	8.02	5.42	30.36	5.00	1.34	0.43	10.45	
Molokai	1.29	2.64	4.57	6.64	7.91	5.78	8.30	6.08	4.88	52.00	
Nickolson	0.67	0.31	0.44	0.42	1.18	12.90	13.47	15.27	5.41	49.89	
Wagram	7.48	20.70	32.06	21.81	5.84	2.00	1.37	1.59	3.31	3.84	

* USDA Taxonomic Names are: Anthony = torrifluents, Ava = fragiudalf, Chalmers = endoaquolls, Davidson = paleudult, Fanno = haplustalf, Kalkaska = haplargid, Mohave = calciargid, Molokai = eutrotox, Nickolson = fragiudalf, Wagram = paleudult.

Source: From Hendricks, D.M., personal communication, 1997.

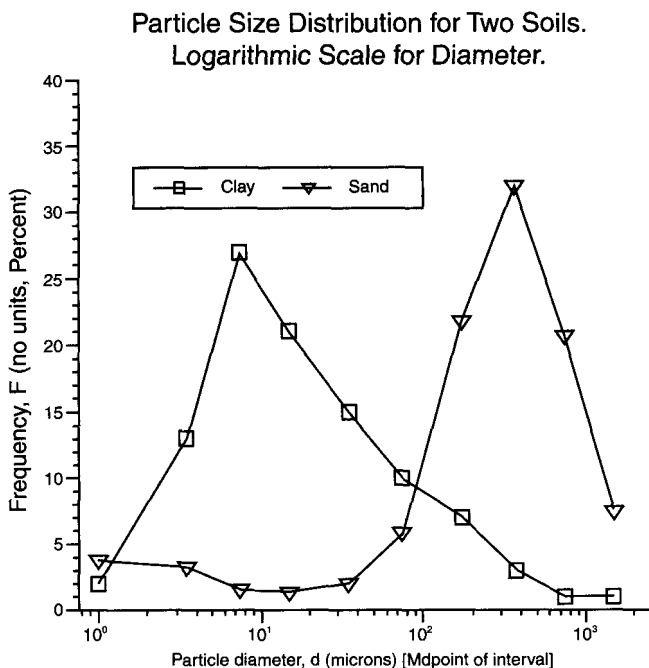


FIGURE 1.2 Frequency graph of particle size distribution.

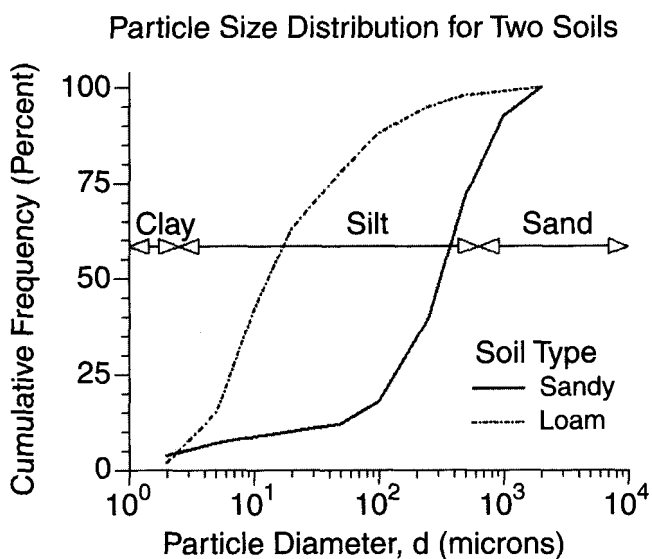


FIGURE 1.3 Cumulative frequency graph of particle size distribution.

1.4.4 SIEVING

The process of sieving is that of placing the particles on a pattern of holes. Small particles may fall through while the sieve retains the larger particles. Either air or water may be the fluid to support the particles as they sort on the sieve. Dry sieving has a lower practical limit of 50 μm , while wet sieving can separate somewhat smaller particle sizes. Sieve holes may be square (using a wire cloth mesh) or round, although square holes are most common. The use of sieves with square openings will not result in measurements equivalent to those using sieves with round holes.

The use of words such as effective or nominal diameters with sieves is in recognition of the imperfect separation that may occur. Placement of a soil sample on a sieve does not result in instantaneous separation. Several factors influence the time to achieve a fixed level of separation. These factors include: sample size, shaking intensity, particle shape, particle size and hole geometry. Since samples vary in their sieving characteristics, it is best to run a trial sample. Errors on a single set of sieves typically are less than 1%, while comparisons between sieves show random errors of about 4%.

Many of the standard sieve sizes correspond to the class limits for USDA soil separates. Surprisingly, no standard sieve is available for the 50 μm cutoff between the sand and silt separates. Consequently, sieving cannot distinguish this class boundary using standard sieves.

1.4.5 SEDIMENTATION

Below 50 μm , sieving is an inefficient and difficult procedure. For soil samples, sedimentation in water is one alternative procedure. A suspension of the dispersed sample settles in water, and at preselected times a measurement is made of the density of particles (mass of particles in a volume of liquid) at a specified depth within the sedimentation cylinder. Variations in the method occur as to the determination of suspension density. In all cases, Stokes' Law is central to the derivation of an equation which relates the time of settling to the size of particle sampled.

Two classic means of determining the density of a suspension exist: hydrometer method and pipet method. In the hydrometer method, the influence of density on a floating object (the hydrometer) is observed. As density decreases (due to settling out of soil particles), the hydrometer sinks. A calibration scale converts the depth of the float (i.e., hydrometer) to the suspension density (expressed as g/L). The pipet method directly removes a sample from the suspension. The concentration of solids is determined by drying the pipeted sample. Nondestructive gamma ray absorption provides a commercially available alternative to both pipet and hydrometer. This method has the advantages of undisturbed and repeated sampling.

The hydrometer method uses higher concentrations of soil and may be less accurate than the pipet method. However, the hydrometer method allows repeated sampling at many points of the distribution (since no sample is withdrawn). Problems exist with the pipet method due to convection currents near the tip of the pipet, effect of sample removal and greater potential for operator error. These problems suggest that the hydrometer method may be preferable in some circumstances.

Stokes' Law for the viscous drag on a sphere is combined with buoyant and gravitational forces to obtain a settling equation. The particle shape is also assumed to be a sphere in evaluating the other forces. Combining forces and solving for v (the particle velocity) gives rise to the settling equation:

$$v = 2r^2g(\rho_p - \rho_w)/9\eta \quad [1.3]$$

where r is the equivalent radius, g is the gravitational acceleration, ρ_w is the fluid density, and η is the viscosity.

The particle velocity is the distance traveled divided by the time, or x/t . The settling equation allows a particle radius to be calculated at a particular x and t , if the particle density and solution viscosity are known. This equation is basic to all gravity sedimentation procedures but requires a number of assumptions.

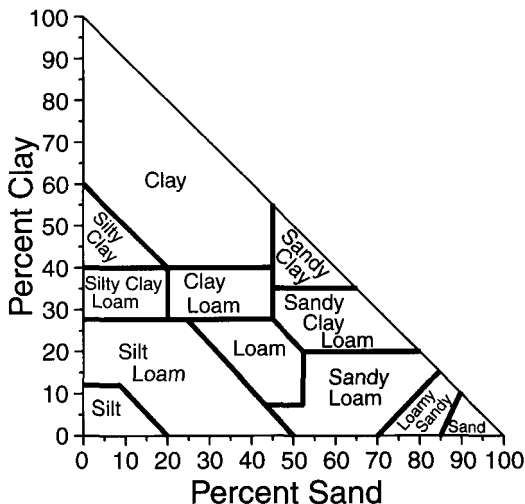


FIGURE 1.4 Alternative texture triangle using clay (%) and sand (%) to determine texture class name. For example, a soil with 50% clay and 30% sand-sized particles would be assigned a textural class name of clay (from the Proceedings of the Soil Science Society of America, 1962, vol. 26, pg. 613. With permission.)

1.4.6 OTHER METHODS

Alternatives exist to fractionation by settling under the influence of gravity in water. Settling can be speeded up, through the use of a centrifuge. Settling can occur in air, called air elutriation. The settling equation for air elutriation is the same as for water except that the velocity is that of the air floating particles upward. Complete avoidance of settling may occur using microscopic methods. Image analysis in conjunction with microscopic methods can determine particle shape or geometry parameters. Other techniques include: conductometric (e.g., Coulter counter) and light scattering.

1.4.7 INTERPRETATION OF RESULTS

A graph can clearly present the elements of a PSD. However, it is desirable to have a more compact means to express the properties of the distribution. This is the origin of soil textural class which summarizes the particle size properties in a single phrase. The phrase chosen depends on the relative abundance of sand, silt and clay, irrespective of variations within these size ranges, through the use of a textural triangle (Figure 1.4). Typical textural triangles and the textural class names are given in Gee and Bauder (1986) and Loveland and Whalley (1991). Note that while the names of three soil separates are similar to the names of textural classes, a textural class name does not limit the sizes that may be present. In other words, a clay texture class contains sand, silt and clay separates while the clay separate contains only clay-sized particles.

The use of only the total amount of sand, silt and clay to describe texture results in a bias which is the assumption that all particles within the range of the sand (or silt or clay) size are equivalent. For the clay fraction, this is a particularly misleading assumption and is partly responsible for the low correlations frequently observed when using clay in regression analysis. Two soils with the same % clay may differ in the amounts of fine clay versus coarse clay. Additional differences in the shape and mineralogy of particles can also cause variations in soil behavior.

A more interesting approach determines the moments of the size distribution. This gives a quantitative measure of the mean, spread and asymmetry of a PSD. More importantly, it increases the power of correlative studies relating texture and any other physical, chemical or biological factors of interest.

Other measures exist in the engineering literature and elsewhere to describe a PSD. One approach defines the grain diameter (D_n) at which $n\%$ passes through a sieve. Therefore, D_{40} represents that size for which 40% of the particles are smaller and 60% are larger. Various combinations of these D_n characterize the distribution. One example is the uniformity coefficient:

$$C_u = D_{60}/D_{10} \quad [1.4]$$

The uniformity coefficient provides information as to how narrow the distribution is, with 1. being the minimum when only a single size is present. The second ratio presented is the coefficient of gradation (C_g):

$$C_g = (D_{30})^2/D_{60}D_{10} \quad [1.5]$$

Indirect descriptions of the particle size distribution are possible by using an equation or model. These models contain parameters which in turn characterize the distribution. The problem in using these models is first to determine the parameters, second, to determine the appropriateness of the model, and third, to interpret the parameters.

One model is of particular interest: the power function. The use of a power function suggests an underlying fractal process. Where this applies, the exponent (δ) relates to the fractal dimension (n) as $\delta = 3 - n$, and n is between 0 and 3. Tyler and Wheatcraft (1992) apply this technique to several materials and report n values between 2 and 3. The log normal model and parameters have also been useful. Campbell (1985) uses these to estimate the water-holding properties of soil.

EXAMPLE 1.3 Particle size analysis by sedimentation. How long does a particle of diameter 0.1 mm take to fall 10 cm in water at 25°C?

Use the formula: $v = 2(\rho_p - \rho_f)gr^2/9\eta$

Assume rapid steady state or: $v = x/t$ Solve for t :

$$t = 9\eta x/2(\rho_p - \rho_f)gr^2$$

Substitute for variables and assume: $\rho_p = 2.65 \text{ Mg/m}^3$

Viscosity and density from CRC Handbook of Chemistry and Physics.

$$\eta = .8904 \text{ cp at } 25^\circ\text{C} \quad \text{and} \quad \rho_f = .997 \text{ Mg/m}^3 \text{ at } 25^\circ\text{C}$$

$$g = 9.80 \text{ m/sec}^2 \quad \text{and} \quad r = .05 \text{ mm}$$

$$t = 9 (.8904 \text{ cp}) (.10 \text{ m})/2(2.65 \text{ Mg/m}^3 - 1.00 \text{ Mg/m}^3)$$

$$* (9.80 \text{ m/sec}^2) (.00005 \text{ m})^2$$

Convert units (this may be done prior to substitution):

$$t = 9 (.890 \text{ g/m-sec})(\text{Mg}/10^6 \text{ g})(.10 \text{ m})/2(2.65 \text{ Mg/m}^3 - 1.00 \text{ Mg/m}^3)$$

$$* (9.80 \text{ m/sec}^2) (.00005 \text{ m})^2$$

Calculate and cancel units to obtain: $t = 9.91 \text{ sec}$

TABLE 1.3
Ranges of specific surfaces of clay minerals,
selected soil components and whole soils
compiled from a variety of sources.

Component	Specific Surface (m ² /g)
Kaolinite	15–20
Illites	80–100
Bentonite	115–260
Montmorillonite	280–500
Organic Matter	560–800
Calcite	.047
Crystalline Iron Oxides	116–184
Amorphous Iron Oxides	305–412
Soils	<10
Sands	5–20
Sandy loams and silt loams	15–40
Clay loams	>25
Clay	>25

1.5 SPECIFIC SURFACE AREA

1.5.1 DEFINITIONS

The surface area of the individual particles is an important factor in nutrient or pesticide adsorption, water absorption, soil strength and soil transport properties. The surface area of a soil has contributions from primary particles, amorphous mineral coatings and organic matter. These individual contributions may overlap and cancel. Further, the surface area of some expanding minerals may change with the water content and chemical composition of the soil solution.

The surface area is an extensive quantity (i.e., depends on how much soil is present). A more satisfying alternative is the introduction of an intensive quantity, the specific surface, which is either the surface area per mass (S_m) or per volume (S_v). The specific surface per volume changes with soil compaction.

1.5.2 TYPICAL VALUES — SEPARATES, WHOLE SOIL (S_m)

Table 1.3 shows typical values for specific minerals as well as values for whole soils. Amorphous materials and soil organic matter can greatly affect soil values. The whole soil values include the effects of particle size distribution as well as typical soil organic matter and amorphous mineral levels for temperate zone soils.

1.5.3 METHODS

Determination of surface area occurs in a number of ways. Direct measurements usually rely on the adsorption of either a gas (typically nitrogen or argon under high vacuum) or a liquid (historically ethylene glycol; but more recently, ethylene glycol monoethylether-EGME or water).

Either a multi- or monomolecular film is deposited or a gas adsorption isotherm is determined. A critical point in the use of all these procedures is the means by which we ensure a multi- or monomolecular layer. Control of the adsorbed phase occurs through regulation of the gas phase. More commonly for liquid adsorption, the use of a desiccator helps to fix the total pressure and partial pressure of the adsorbed component. In the EGME method a mixture of EGME and CaCl_2 regulates the vapor pressure of EGME. In water absorption a separate saturated lithium nitrate solution regulates the relative humidity. The lithium nitrate and EGME methods are convenient procedures for soils; however, both methods are sensitive to variations in temperature. Values for specific surface vary with the method used.

Another direct method examines photomicrographs of primary particles. The classical approach determines the probability of a needle (randomly placed on the photo) either falling within a pore or intersecting the pore edge. Current image analysis instrumentation allows this evaluation (through the particle size distribution) as part of many software packages. Unfortunately, most particle size distributions are not detailed enough in the smallest size range to accurately estimate specific surface.

1.5.4 MODELS

Models are used to describe how particle size influences specific surface and the general relation of particle geometry to specific surface. Starting with the specific surface $S_m = A/\rho_p V$, where A is the surface area of solids and V is the volume of solids, A and V are expressed in terms of the shape factors C_2 and C_3 with a characteristic dimension L :

$$S_m = C_2/\rho_p C_3 L \quad [1.6]$$

This relation states that specific surface is the reciprocal of the characteristic length. The shape factors also influence surface area. Using Table 1.1, for spheres: $S_m = 6/\rho_p d$; while for a disc or flat plate: $S_m = [4 + 2(d/h)]/\rho_p d$. A large d/h ratio corresponds to a flat plate, while a small d/h ratio corresponds to a prism or needle-like shape. A ratio of $d/h = 1$ is identical to the result for spheres (Figure 1.5). The specific surface of fine clays may be one or more orders of magnitude greater than for coarse clays.

EXAMPLE 1.4 Specific surface area from particle shape factors. Derive a relation for the specific surface area of rod-shaped particles.

Start with the specific surface $S_m = A/\rho_p V$, where A is the surface area of solids and V is the volume of solids. Then express A and V in terms of the shape factors C_2 and C_3 from Table 1.1 with a characteristic dimension L :

$$S_m = C_2 L^2 / \rho_p C_3 L^3$$

$$S_m = C_2 / \rho_p C_3 L$$

$$S_m = [2 + (4h/d)] / \rho_p h$$

where h is the rod length (and characteristic length) and d rod edge length.

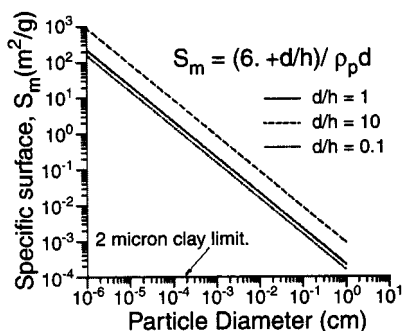


FIGURE 1.5 Effect of particle diameter on specific surface for materials of a single size.

1.6 BULK DENSITY (ρ_B) AND POROSITY (φ)

1.6.1 DEFINITIONS

To quantify the state of compaction and the amount of pore space in a soil, the volume and mass of solids which are extensive quantities are related by the intensive term, bulk density (ρ_B):

$$\rho_B = (\text{mass of solids}) / (\text{volume of solids and voids}) \quad [1.7]$$

The voids are the pore spaces that may hold either air, water or other liquids.

A related term to density is specific gravity. Specific gravity is defined as the ratio of the density (particle, bulk, fluid or any other density) to the density of water at 4°C and standard pressure. The specific gravity is dimensionless and the reference density of water is exactly 1.00 Mg/m³.

1.6.2 TYPICAL VALUES

The bulk density is a key physical property of any porous material and changes in response to disturbance or soil management practices. Yet there is a limit to any modification of bulk density. The particle size distribution along with packing controls the range of possible values.

Bulk density varies with the packing of the soil particles. Typically, sands pack more closely and values range from 1.4 to 1.9 Mg/m³. Clays tend to bridge and cannot pack as tightly, giving values from 0.9 to 1.4 Mg/m³. Textures between sands and clay vary in their bulk density accordingly. The wide range in bulk densities for a particular texture indicates that other factors (such as organic matter and compactive history) have an important influence on this property.

Field determinations of bulk density have relatively low precision (typically $\pm .05$ Mg/m³), which means that only about 10–20 different states of bulk density can be distinguished. An increase in measurement precision is inadequate because sampling bias and natural variability are similar in magnitude to typical measurement errors. Bulk density typically has coefficients of variations in the range of 10–40%.

Bulk density is highly dependent on soil conditions at the time of sampling. Changes in soil swelling due to changes in water content alters the bulk density. Thus, comparisons of bulk density must control or compensate for water content where swelling is significant. Other factors such as traffic patterns can also influence the bulk density. Consequently, a determination of bulk density may require the use of the above definition and the specification of the conditions at the time of sampling.

If the bulk density of a soil is fixed, then the relative amount of pore space is also fixed. To make this concept more precise, a term is needed to describe the amount of pore space (or voids). With pore space, volume rather than mass dimensions are more appropriate. Porosity as an intensive quantity is defined as:

$$\text{Porosity} = \phi = \text{volume of voids/volume soil} \quad [1.8]$$

Alternately, it can be shown that this definition is equivalent to the following (if the particle density includes organic matter):

$$\phi = 1. - (\rho_B/\rho_P) \quad [1.9]$$

This relation is the result of definitions and is not empirical. Using typical bulk densities, the total porosity of sandy soils is less than that of finer textured soils. It implies that for every value of bulk density, a given soil has only one possible value of the porosity. However, it is not true that a soil has only one possible value of the bulk density.

1.6.3 RELATED TERMS

A number of other expressions which characterize the amount of air in a soil are listed with their names, symbols and definitions:

$$\text{Air Filled Porosity} = \phi_a = \text{volume air/volume soil} = \phi - \theta_v, \quad [1.10]$$

where θ_v is the volumetric water content of a soil and defined in a later chapter.

$$\text{Void Ratio} = \text{volume of voids/volume of soils} = \phi/(1 - \phi) \quad [1.11]$$

$$\text{Wet Density} = \rho_M = (\text{mass of solids} + \text{water})/\text{volume of soil} \quad [1.12]$$

Note that “wet” in the last expression refers to the inclusion of water in the numerator and not to how much water is present. It is possible to calculate the wet density of a dry soil and to convert between bulk density and wet density using the relation:

$$\rho_B = \rho_M - \theta_v \rho_w \quad [1.13]$$

Wet density is not a preferred means of expressing the packing density of a soil.

1.6.4 FURTHER USES FOR BULK DENSITY

Another use for bulk density is the conversion of any gravimetric quantity (i.e., an intensive quantity that we express on a per gram basis) to a volumetric basis. For example, in Section 1.5 there are two kinds of specific surface: S_v and S_m . These are surface areas per quantity of volume or mass. The explicit relation between the two is:

$$S_v = S_m \rho_B \quad [1.14]$$

Some researchers try relating bulk density to factors like root penetration, soil strength and compaction (Table 1.4; SCS, 1981). These attempts generally meet with mixed success.

TABLE 1.4
Approximate bulk densities that restrict root penetration.

Texture	Critical Bulk Density for Soil Resistance	
	High (Mg/m ³)	Low (Mg/m ³)
Sandy	1.85	1.60
Coarse-loamy	1.80	1.40
Fine-loamy	1.70	1.40
Coarse, Fine-Silty	1.60	1.30
Clayey	(Depends on both clay % and structure)	

1.6.5 METHODS

Typically, soil bulk density is determined by inserting a ring into the soil. The ring is of known volume and upon extraction the soil core within the ring is dried to determine mass of solids and water present at the time of sampling. The major difficulties are first, the presence of stones or organic matter (possibly alive) and second, the compaction of the core while sampling may bias the volume. The effect of stones and compaction can be minimized by using a larger sampling ring.

There will be sites where a ring technique is not feasible. For example, coarse-textured soils may not remain in the ring or it may be difficult to drive the sampling ring to the desired depth. Further, since soil compressibility depends on water content, there will be times when sampling a particular site may show bias because the soil is too wet. The presence of live or dead roots also poses a problem, particularly near the soil surface and in soils managed with reduced tillage.

An alternative field method relies on hand-excavating soil. This results in an irregularly shaped hole whose volume we must determine. The hole is filled with a measurable volume (sand, an air-filled balloon or a water-filled balloon) which allows accurate calibration. Gamma ray attenuation methods have been developed for field use. Details can be found in references by Campbell and Henshall (1991) and Culley (1993).

Laboratory columns allow for greater precision in the direct determination of bulk density. All packed columns show systematic variations with depth which depend on the packing technique used. One alternative laboratory procedure uses gamma ray attenuation. Other techniques that may have application include computer assisted tomography (CT) and sensing of soil dielectric properties.

REFERENCES

- Anderson, S.H., R.L. Peyton and C.J. Gantzer. 1990. Evaluation of constructed and natural soil macropores using x-ray computed tomography. *Geoderma*, 46:13–29.
- Baver, L.D., W.H. Gardner and W.R. Gardner. 1972. *Soil Physics*, 4th ed., John Wiley and Sons, New York.
- Blake, G.R. and K.H. Hartge. 1986. *Methods of Soil Analysis*, pt. 1, chap. 14, A. Klute (ed.), ASA, Madison, WI.
- Borggaard, O.K. 1982. The influence of iron oxides on the surface area of soil, *J. Soil Sci.*, 33:443–449.
- Bower, C.A. and F.B. Gschwend. 1952. Ethylene glycol retention by soils as a measure of surface area and interlayer swelling, *SSSA Proc.*, 16:342–345.
- Campbell, D.J. and J.K. Henshall. 1991. Bulk density in soil analysis physical methods, chap. 7, K.A. Smith and C.E. Mullins (eds.), M. Dekker, Inc., New York.
- Campbell, G.S. 1985. *Soil physics with basic*, Elsevier, Amsterdam.
- Carter, D.L., M.M. Mortland and W.D. Kemper. 1986. Specific surface, in *Methods of Soil Analysis*, pt. 1, chap. 16, Physical and mineralogical methods, ASA, Madison, WI.

- Culley, J.L.B. 1993. Density and compressibility in soil sampling and methods of analysis, chap. 50, M.R. Carter (ed.), *Canadian Society of Soil Sci. Lewis Publ.*, Boca Raton, FL, p. 823.
- Davies, R. 1984. Particle size measurement: experimental techniques, in *Handbook of Powder Science and Technology*, chap. 2, M.E. Fayel and L. Otten (ed.), Van Nostrand Reinhold Co., New York.
- Egashira, K. and J. Matsumoto. 1981. Evaluation of the axial ratio of soil clays from gray lowland soils based on viscosity measurements, *Soil Sci. Plant Nutr.*, 27:273–279.
- Einstein, A. 1906. A new determination of molecular dimensions, *Annalen der Physik.*, 19:289–306, translated in: *Investigations on the Theory of Brownian Movement*, Dover Press.
- Elder, J.P. 1979. Density measurements by the mechanical oscillator, in *Methods in Enzymology*, vol. 61, Academic Press, 12–25.
- Elghamry, W. and M. Elashkar. 1962. Simplified textural classification triangles, *Soil Sci. Soc. Am. Proc.*, 26:612–613.
- Folk, R.L. 1966. A review of grain-size parameters, *Sedimentology*, 6:73–93.
- Gee, G.W. and J.W. Bauder. 1979. Particle size analysis by hydrometer: a simplified method for routine textural analysis and a sensitivity test of measurement parameters, *Soil Sci. Soc. Am. Proc.*, 43:1004–1007.
- Gee, G.W. and J.W. Bauder. 1986. Particle size analysis, in *Methods of Soil Analysis*, pt. 1, chap. 15, Second ed., A. Klute (ed.), ASA, Madison, WI. 383–411.
- Heywood, H. 1947. Symposium on particle size analysis, *Trans. Inst. Chem. Eng.*, 22:214.
- Jensen, E. and H.M. Hansen. 1961. An elutriator for particle-size fractionation in the sub-sieve range, *Soil Sci.*, 92:94–99.
- Kahn, A. 1959. Studies on the size and shape of clay particles in aqueous suspensions, *Clays Clay Miner.*, 6:220–236.
- Karsten, J.H.M. and W.A.G. Kotze. 1984. Soil particle size analysis with the gamma attenuation technique, *Commun. Soil Sci., Plant Anal.*, 15:731–739.
- Klein, C. and C.S. Hurlbut, Jr. 1985. *Manual of Mineralogy* (after James D. Dana), 20th edition, Wiley, New York.
- Loveland, P.J. and W.R. Whalley. 1991. Soil analysis physical methods, in K.A. Smith and C.E. Mullins, (eds.), M. Dekker, Inc., New York, 620.
- Nelson, R.A. and S.B. Hendricks. 1943. Specific surface of some clay minerals, soils, and soil colloids, *Soil Sci.*, 56:285–296.
- Orchiston, H.D. 1955. Adsorption of water vapor, III. Homoionic montmorillonites at 25°C, *Soil Sci.*, 79:71–78.
- Santo, L.T. and G.Y. Tsuji. 1977. Soil bulk density and water content measurements by gamma-ray attenuation techniques, *Tech. Bull. 98*, Hawaii Agric. Expt. Stn.
- Scheidegger, A.E. 1960. *The Physics of Flow through Porous Media*, revised edition, University of Toronto Press.
- SCS. 1981. Tables for discussion of estimation of bulk density, Approximation 3, In-house report of NSSL, R.B. Grossman, Lincoln, NE.
- Sheldrick, B.H. and C. Wang. 1993. Particle size distribution in soil sampling and methods of analysis, M.R. Carter, (ed.), chap. 47, *Canadian Society of Soil Sci. Lewis Publ.*, Boca Raton, FL, 823.
- Shirazi, M.A. and L. Boersma. 1984. A unifying quantitative analysis of soil texture, *Soil Sci. Soc. Am. J.*, 48:142–147.
- Streeter, V.L. and E.B. Wylie. 1975. Fluid mechanics, sixth edition, McGraw-Hill Inc., New York.
- Suarez, D.L. and J.D. Wood. 1984. Simultaneous determination of calcite surface area and content in soils, *Soil Sci. Soc. Am. J.*, 48:1232–1235.
- Svedberg, T. and J.B. Nichols. 1923. Determination of size and distribution of size of particles by centrifugal methods, *J. Am. Chem. Soc.*, 45:2910–2917.
- Tanner, C.B. and S.J. Bourget. 1952. Particle-shape discrimination of round- and square-holed sieves, *Soil Sci. Soc. Am. Proc.*, 16:88.
- Tyler, S.W. and S.W. Wheatcraft. 1992. Fractal scaling of soil particle-size distributions: analysis and limitations, *Soil Sci. Soc. Am. J.*, 56:362–369.
- Uhland, R.E. 1949. Physical properties of soils as modified by crops and management, *SSSA Proc.*, 14:361–366.