Technical Note 2.
Soil Particle and Bulk Density

2.0 Preamble

Particle and bulk density are two fundamental physical properties of soil. Particle density reflects the mineral and organic composition of the soil solid phase, with most mineral soils approximating 2.65 g/cm³. Particle density is typically measured in a laboratory with special apparatus. Bulk density can be measured in the field or laboratory. Bulk density has many practical applications, for example:

- Calculating the mass of soil per unit volume, area, and depth
- Calculating total water storage capacity per volume of soil
- Calculating plant-essential nutrients per unit area and depth of soil
- Calculating soil porosity (if particle density is known)
- Evaluating the significance of soil compaction

If an inveterate dirt hog were to go into the field and extract carefully, a 1 x 1 meter block of soil, full inspection would reveal that it is made up of four components:

- **Mineral Solids**: Particles ranging in size from gravel and sand, visible to the naked eye, to silt particles requiring a hand lens, to clay particles visible only under electron microscopy. Mineral particles form the structural framework of the soil.
- **Pores**: The open spaces between particles containing air, water, or both; their size distribution governs water retention, drainage, and gas exchange.
- **Water**: The liquid phase containing dissolved minerals, nutrients, and organic compounds that occupies pore spaces to varying degrees.
- **Organic matter**: The decayed or partly decayed remains of once-living organisms. Organic matter accumulates in the topsoil, imparting a dark brownish color, and influences soil structure and water retention.

In this technical note, the treatment of soil particles and bulk density rests on the first two components: mineral solids and pore space. Soil water is not considered when measuring particle or bulk (dry) density. Organic matter is also ignored. However, water and organic matter content strongly influence the degree of ***soil compactibility***.

The relationship between bulk density and soil compaction is a perennial topic about which much has been written. Here, we examine several soil properties that influence compactibility and the practical consequences, both good and bad, of soil compaction.

2.1 Soil Particle Density

Density is a universal property of all substances—everything that has mass and occupies space. This applies to solids, liquids, gases, and plasmas. In principle, density is defined as mass per unit volume:

$$\rho = \frac{m}{V}$$

where ρ is the density (symbolized by the lower-case Greek letter rho), m is the mass, and V is the volume. Units for density may be grams per cubic centimeter (g/cm³: CGS metric), kilograms per cubic meter (kg/m³: MKS metric), or pounds per cubic foot (lb/ft³: English).

Physical materials have different densities depending on their chemical structure and degree of purity. To simplify comparison of density across substances and units of measurement, a dimensionless ratio, "relative density" or "specific gravity," is often used instead. Here, the ratio of a substance is always compared to that of a standard substance, usually pure water at 4 °C, which has a density of 1 g/cm³. The density of water varies slightly with temperature. However, for many practical applications, a density of 1 g/cm³ can be assumed.

Particle density, ρ_s, is a physical property of soil, defined as the mass per unit volume of solid particles, reported in grams per cubic centimeter (g/cm³: metric). Particle density depends upon the chemical composition and crystal structure of the mineral particles and is not affected by pore space. As such, particle density is independent of particle size and soil arrangement.

Mineral particles (sand, silt, clay) have higher densities compared to organic matter. Typical particle densities range from 2.6-2.75 g/cm³, with 2.65 g/cm³ often used as an "average" value for expressions of density. The particle density range is narrow because quartz, feldspar, mica, and clay minerals make up the bulk of mineral soils, and their densities fall within this range.

By comparison, water, pine wood, concrete, mild steel, and lead metal have densities of 1.0, 0.7, 2.4, 7.8, and 11.3 g/cm³ (Table 1).

Table 1. Density of various substances.

	g/cm³	lb/ft³	kg/m³
Water	1.0	62.4	1,000
Pine wood	0.7	44	700
Concrete	2.4	150	2,400
Steel, mild	7.8	480	7,700
Lead metal	11.3	706	11,300

Organic matter has a lower density than mineral particles, ranging from 0.9 to 1.3 g/cm³. The topsoil layer typically has a higher organic matter content and lower bulk density than the subsoil. Since bulk density accounts for total soil volume, solids and pore space combined, it will always be lower than the particle density of the mineral or organic constituents themselves.

A volumetric glass flask, called a pycnometer or specific-gravity flask, is used to measure soil particle density. The flask is fitted with a ground-glass stopper perforated lengthwise with a capillary channel (Figure 1a and 1b). Sometimes a thermometer is included in the apparatus, but it's not an integral component. The capillary tubing is immersed in the mineral particle-water mixture via the capillary channel. Entrapped air is removed by applying vacuum pressure through the capillary tubing. Specific gravity is calculated using a mathematical formula covered by Blake and Hartge (1986) and ASTM D854.

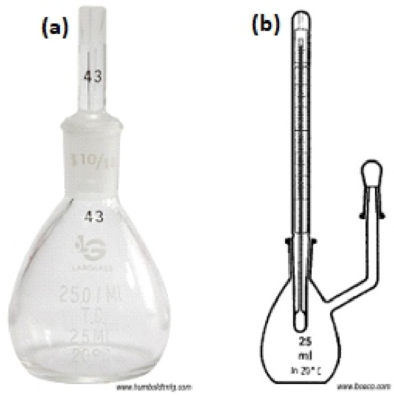


Figure 1 Pycnometer flasks for determining soil particle density (**a**) without and (**b**) with a thermometer.

2.2 Soil Bulk Density

Bulk density is the mass of dry (moisture-free) soil per unit volume of soil in its undisturbed condition. The formula for bulk density is:

$$\rho_b = \frac{\text{soil mass oven-dry}}{\text{soil volume}}$$

where ρ_b is the bulk density (symbolized by the lower-case Greek letter rho subscripted with the lower-case letter b). Where bulk density is computed on a moisture-free (i.e., oven-dry) basis, the amount of water in the soil is ignored[1]. Note that bulk density is obtained by dividing by the total "bulk" sample volume, including all particle solids and pore space, by the oven-dry soil mass. In contrast, density is obtained by dividing the mass of a substance by its volume. In geotechnical practice, the term *dry density* is often used instead of bulk density to distinguish from the in-situ moist condition.

Compared to particle density, bulk density accounts for the total pore space in a soil body. Bulk density can provide insight to other soil properties such as porosity, water storage, fluid permeability, and compaction. The bulk density of mineral soil usually ranges from 1.0 to 1.8 g/cm^3, varying primarily with:

- **Pore space** An average, well-aggregated soil has approximately 50 percent pore space. In general, soils with less pore space are more compacted; soils with more pore space are less compacted. Soil with less than 45% pore space by volume is not optimal for agricultural use.
- **Texture** Fine-textured soils such as silt loam, clay, and clay loam generally have lower bulk density than sandy soils. Bulk density levels for cultivated, fine-textured soils range from 1.0 - 1.6 g/cm^3, while sandy soils have bulk densities greater than 1.4 g/cm^3 (Table 2). These differences are primarily due to greater soil aggregation and structure in fine-textured soils.
- **Organic matter** The organic matter content of soil affects bulk density: generally, bulk density decreases as organic matter content increases, and vice versa. Organic soils commonly have a bulk density < 1.0 g/cm^3. Sandy soils tend to have lower organic matter and a granular structure. As a result, they often have higher bulk densities than fine-textured soils.
- **Soil depth** Bulk density tends to increase with depth in the soil profile. This is due to (a) lower organic matter content in the subsoil; (b) compaction due to the weight of the overburden material; (c) less aggregation, and (d) less root, earthworm, and meso-fauna activity.

[1]Bulk density can be calculated on a dry or wet basis. Soil scientists usually report dry density, though this is not always explicitly stated. Wet density is measured when information on the total mass of soil and water per unit volume is needed, for example, when evaluating the bearing capacity of structural members under the weight of soil at a given water content. Because wet bulk density changes with water content, dry bulk density provides a more stable and reproducible characterization of soil structure.

Table 2. General relationships among texture, particle and bulk density, and porosity of soils.

Soil Type	Particle density (g/cm^3)	Bulk density (g/cm^3)	Porosity (cm^3/cm^3)
Mineral soils, plowed horizons			
fine-medium texture	2.60	1.0-1.6	0.69-0.46
coarse texture	2.60	1.4-1.8	0.46-0.35
Subsoil and parent materials	2.65	1.5-1.8	0.43-0.32
Grassland and woodland,			
surface horizon	2.4	0.8-1.2	0.67-0.50
Peats	1.4	0.1-0.3	0.93-0.79

Source: Rowell (1994) and McCarty et al. (2016)

Bulk density is determined by measuring the mass of soil in a container of known volume, typically a cylinder. The volume of the cylinders inside is equal to the soil volume. After pushing the soil out of the cylinder, it is dried in an oven at 105° C until it reaches a constant weight. Bulk density is calculated by dividing the oven-dry mass by the cylinder volume. It is not necessary to extract a uniform-shaped soil sample for bulk density. Any volume of soil can be extracted, provided the excavation volume can be determined. Clods can be coated with liquid paraffin and dipped into water to measure water displacement (Archimedes' principle) and, in turn, calculate volume. One piece of equipment for measuring in-situ bulk density consists of a smooth-walled, single- or double-cylinder sampler equipped with a drive head and hammer assembly (Figure 2).

2.3 Soil Porosity

Soil porosity, *pore space*, or *void space*, is that portion of the soil volume not occupied by solid particles, either mineral or organic. An "ideal" soil for crop production would contain 25 percent water and air by volume, and 50 percent consisting of solid particles (Figure 3).

The solid portion is made up of the minerals sand, silt, and clay, and a small fraction (usually < 4%) of organic matter. Under field conditions, air and water occupy the pore space between the mineral particles at all times. The size of individual pores depends on the shape and size of primary particles and how tightly they are packed together. When solid particles are in close contact with each other, porosity is predicted to be low, and bulk density high. When particles are arranged in porous aggregates, as so often in medium-textured soils high in organic matter, the volume of pore space is predicted to be high and bulk density low.

The formula for calculating the percentage of the total volume of soil occupied by pore space is given by:

$$f_t = \left(1 - \frac{\text{bulk density, } \rho_b}{\text{particle density, } \rho_s}\right) \times 100$$

where f_t is the fractional volume of soil occupied by pore space (air- or water-filled) expressed as a percentage of the total volume of soil[2] (symbolized by the lower-case letter *t*).

Porosity (f_t)[3] is calculated from two density measurements — particle density (Section 2.1) and bulk density (Section 2.2) — using the formula $f_t = 1 - (\rho_b/\rho_s)$. For most mineral soils, porosity values generally range from 30%–60%. Sandy soils typically vary in total pore space from 30%–50%, whereas fine-textured soils vary from 40%–60%. Notable exceptions include andic soils (Andisols) derived from volcanic materials, which can exceed 70% porosity, and heavily compacted soils, which may fall below 30%.

It is important to understand that bulk density differences between clay and sandy soils are primarily explained by differences in total porosity rather than particle density. Clay soils have many small pores, giving high total porosity and correspondingly low bulk density, despite having particle densities similar to those of sandy soils. Sandy soils have larger but fewer pores per unit volume, resulting in lower total porosity and higher bulk density. The particle densities of the dominant minerals in both sand and clay fractions — quartz, feldspar, and phyllosilicate clay minerals — fall within a relatively narrow range of approximately 2.2–2.9 g/cm³, so particle density alone does not explain bulk density differences between textural classes.

Where bulk density is notably lower than expected from mineralogy alone, the presence of organic matter is often a contributing factor, as organic matter has a particle density of approximately 1.3–1.7 g/cm³, substantially lower than mineral constituents. Andic soils developed from volcanic parent materials present a notable exception at the low end, as allophane and other short-range-order minerals have particle densities of 1.9–2.1 g/cm³ and high porosity, resulting in unusually low bulk densities, often below 0.9 g/cm³.

At the high end, highly weathered tropical soils (Oxisols and related orders) dominated by iron and aluminum oxides such as goethite, hematite, and gibbsite can have particle densities ranging from 3.5–5.3 g/cm³, potentially elevating particle density independently of porosity effects, though their characteristically stable microaggregates often maintain higher porosity than their mineral composition alone would suggest.

A worked example calculating particle and bulk density, and soil porosity follows.

Example A cylindrical soil sample was taken for determination of bulk density. The cylinder's inside diameter was 4.12 cm, and its height 10.16 cm, both measured with a caliper. The soil in the cylinder weighed 247.60 grams (g) in its undisturbed moist condition and 218.90 g when oven-dried. After grinding and passing through a sieve with 2 millimeter diameter openings, a 25.00 g subsample was poured into a beaker with 100 milliliters (mL) of distilled water where the soil-water volume was 109 mL. Calculate the bulk density (ρ_b), particle density (ρ_s), and percent pore space (f_t). Assume 1 mL water = 1 cm³ water, i.e., density of 1.0.

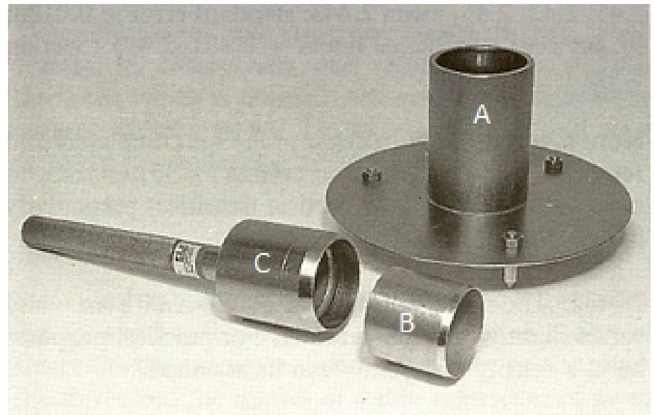


Figure 2 Equipment for measuring soil bulk density in the field. The guide plate and cylinder (A) is placed on the soil surface. The sampling cylinder, or "Shelby" drive tube (B), is machined to fit inside the drive hammer (C), The drive hammer and tube are then inserted into the guide cylinder and hammered into the soil by dropping a sliding weight attached to the stem of the drive hammer (not pictured). The drive hammer includes some headspace allowing soil to extend beyond the drive tube, avoiding sample compaction. (*ELE International/Soiltest*)

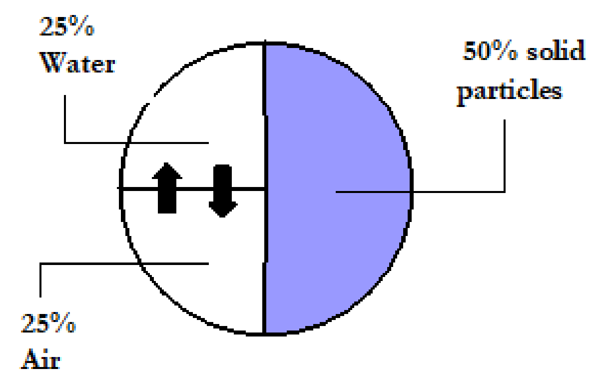


Figure 3 "Ideal" mineral soil for crop production consisting of 25% air and water by volume, and 50% solid particles (minerals and <4% fraction of organic matter). Although the percentages vary, every soil has the same four components: mineral and organic matter, air, and water.

[2]Void ratio is parallel index of porosity employed by geotechnical engineers. Void ratio is calculated by dividing the volume of voids (pore space), by the volume of solids (soil particles). Void ratio relates the volume of pores to the volume of solids. As such void ratio is independent of total volume.

[3]The symbol n is also used as a variable or notation for porosity by some disciplines.

$$\text{cylinder volume } (V_t) = \pi r^2 h$$
$$= (3.14)(2.06 \text{ cm}^2)(10.16 \text{ cm})$$
$$= 135.45 \text{ cm}^3$$

$$\text{volume of water displaced by soil } (V_s)$$
$$= 109.5 \text{ mL} - 100 \text{ mL}$$
$$= 9.5 \text{ mL or } 9.5 \text{ cm}^3$$

$$\text{particle density } (\rho_s) = \frac{m_s}{V_s}$$
$$= \frac{25.00 \text{ g}}{9.5 \text{ cm}^3}$$
$$= 2.63 \text{ g/cm}^3$$

$$\text{bulk density } (\rho_b) = \frac{m_s}{V_t}$$
$$= \frac{218.90 \text{ g}}{135.45 \text{ cm}^3}$$
$$= 1.62 \text{ g/cm}^3$$

$$\text{total porosity } (f_t) = \left(1 - \frac{\text{bulk density}, \rho_b}{\text{particle density}, \rho_s}\right) \times 100$$
$$= \left(1 - \frac{1.62}{2.63}\right) \times 100$$
$$= 38.4\%$$

In the example above, particle density was calculated using the displacement method. The "average" particle density of 2.65 g/cm^3 is used in many applications.

With a bit more finagling, we can calculate the percentage of the total pore space occupied by water and air. To calculate water-filled porosity, the volumetric water content (percentage water by volume) must be determined. Volumetric water content is the percentage water content by mass in the sample, multiplied by the sample's bulk density. Mass soil water content is determined as follows:

$$\text{water content } (\theta_g) = \frac{\text{wet soil mass} - \text{dry soil mass}}{\text{dry soil mass}}$$
$$= \frac{247.60 \text{ g} - 218.90 \text{ g}}{218.90 \text{ g}}$$
$$= \frac{28.7 \text{ g}}{218.9 \text{ g}}$$
$$= 0.13 \text{ g } H_2O / \text{g soil } (13\% \text{ by mass})$$

Soil water content, θ_g (symbolized by the lower-case Greek letter theta subscripted by the lower-case letter g), is the mass of water per unit mass of solid particles. For the sample above, each gram of soil contains 0.13 grams of water. The subscript "g" stands for "gravimetric", which means that the water content in the sample was determined by mass difference (in this case, loss of water mass during oven-drying).

The water-filled porosity is determined by:

$$\text{water-filled porosity } (f_w) = \frac{\%\text{ water by volume}, \theta_v}{\text{total porosity}, f_t} \times 100$$
$$= \frac{21\%}{38.4\%} \times 100$$
$$= 55\%$$

The following table summarizes the soil properties determined for our sample.

Property	Value
Particle density	2.63 g/cm^3
Bulk density	1.62 g/cm^3
Total porosity	38.4%
Water-filled porosity	55%
Gravimetric water content	0.13 g/g
Volumetric water content	0.21 cm^3/cm^3

What does this information reveal about the soil?

Particle density of 2.63 g/cm^3 indicates that the sample is a mineral soil with relatively low organic matter content (Table 2). The percentages of sand, silt, and clay are unknown; however, particle density suggests the sample is higher in coarse-texture sand than in fine-texture silt and clay.

From Table 2, a soil with a bulk density of 1.62 g/cm^3 and 38.4% total porosity also suggests that, along with particle density, the soil is coarse-textured. Total porosity includes large and small pores. As such, it tells us nothing about the distribution of pore sizes in the soil. Water is held in small pores less than about 50 microns (µm) in diameter by adhesive and cohesive (surface tension) forces collectively called *capillary forces*. Capillary pores are the main reservoir of plant available water and determine the soil's water holding capacity. In contrast, water drains from pores larger than about 50 microns (µm) in diameter by gravity, a process called noncapillary or macropore flow.

If we assume the soil water content was at ***field capacity*** when the sample was taken, i.e., all gravitational water drained away from the sample zone, the volume of water-filled pore space should roughly equal the volume of capillary pore space. The calculated volume of water-filled pore space in our sample was 55%. In turn, the percentage of air-filled pore space can be found by subtraction: 100 - 55 = 45%. These figures come close to the "ideal" agricultural soil with air-filled (non-capillary) and water-filled (capillary) pore space occupying about equal volume of the total pore space. However, the reader should bear in mind that the relative amounts of water- and air-filled pore space fluctuate constantly. Rainfall infiltrating the soil drives air out of the pores. As the soil dries, air re-enters the pores, gradually reaching field capacity over 1-3 days, depending on soil texture and internal drainage. For growing plants, pore-size distribution is more important than total porosity.

The depth of water in the soil can also be calculated from the volumetric water content since 1 cm^3 of water also occupies a cube 1 cm x 1 cm x 1 cm (l x w x h) if no soil particles are present. Thus, soil with a volumetric moisture content of 0.21 cm^3 would occupy a layer 1 cm long, 1 cm wide, and 0.21 cm high. A volumetric water content of 21% is therefore equivalent to a depth of 0.21 centimeters of water *per* *centimeter depth of soil*. Our core sample measured 10.16 cm deep, so the total depth of water in the core is 10.16 x .21 = 2.13 cm (0.84 inches). This is more water than predicted by coarse texture alone, so our assumption about the soil reaching field capacity may not be true[4].

The exact mass of soil per hectare can be calculated from bulk density using the following equation:

$$\frac{\text{kilograms soil}}{\text{hectare}} =$$

$$\text{cm depth} \times \frac{10{,}000\ \text{m}^2}{\text{hectare}} \times \frac{1\ \text{m}}{100\ \text{cm}} \times \frac{1.62\ \text{g}}{\text{cm}^3} \times \frac{\text{cm}^3}{1\ \text{g H}_2\text{O}} \times \frac{1{,}000\ \text{kg H}_2\text{O}}{1\ \text{m}^3}$$

This equation tells us that, when calculating the mass of 1 hectare (2.47 acres) of soil, its mass is compared to that of water. Soil with a bulk density of 1.62 g/cm³ has a mass 1.62 times that of an equivalent volume of water, or 62% greater mass per unit volume than water. (the density of pure water at 4°C is given as 1 g /cm³). A cubic meter of water has a mass of 1,000 kilograms (symbol: kg). If soil is 1.62 times as heavy, then its mass is:

- 1,000 × 1.62 = 1,620 kg per cubic meter (kg/m³)
- 1,620 kg × 10,000 cubic meters per hectare (100 m × 100 m area ×1 meter thick) = 16,200,000 kg soil (16,200 metric tons)

Since our core measures 10.16 cm deep, the adjusted mass of soil is:

- 16,200,000 × 0.1016 = 1,645,920 kg (~ 1,468,160 lb/acre, 4 inches deep).

Soil testing laboratories typically assume a sampling depth of 6 to 7 inches. The exact mass of soil per acre is usually calculated based on pounds per acre-furrow slice (lb/AFS). By definition, an acre-furrow slice is 6.67 inches deep. It is the assumed depth that a 16-inch moldboard plow bottom reaches. A *furrow slice* of soil is the layer of soil sliced away and inverted by a moldboard plow. Acre-furrow slice is the assumed mass of one acre of soil 6 inches deep, 1,000 tons (2,000 pounds per ton). *Hectare-furrow slice* is the assumed mass of one hectare (2.47 acres) of soil 15 cm deep, 2,200 metric tons (1,000 kilograms per ton).

In practice, plowing may be deeper or shallower than 6.67 inches. As such, acre-furrow slice is not an exact, scientific unit of measure. However, the term *acre-furrow slice* is well established in agronomic practice where an approximation is needed.

A six-inch thickness of soil with a bulk density 1.47 g/cm³ weighs 1,998,731 lb/AFS (~2,240,377 kg/ha) or approximately 2 million pounds. Because one acre-furrow slice contains approximately 2,000,000 lb of soil, each part per million by mass is equivalent to approximately 2 pounds per acre-furrow slice.

Therefore, we can convert quantitative parts per million (symbol: ppm) of a nutrient to pounds per acre furrow slice (6 inches deep) by multiplying ppm by 2.

For example, an elemental potassium (K) concentration of 150 ppm in the 6" inch (15 cm) plow layer of a soil with a bulk density of 1.47 g/cm³ would have a K content of 150 × 2 = 300 lb/acre (336 kg/ha). Generally, the supply of plant essential nutrients in pounds per acre can be approximated by multiplying ppm by a factor of 2 if the sample depth is 6 inches[5]. Soils with substantially different bulk densities will give different conversion factors when sampled to the same depth. For example, a bulk density of 1.35 g/cm³ gives approximately 1.84 × ppm, and a bulk density of 1.6 g/cm³ gives approximately 2.18 × ppm. Note that plant roots often extend deeper than six inches. Estimating the total supply of plant essential nutrients is unreliable without sampling beneath the surface rooting zone.

2.4 Bulk Density and Compaction

Bulk density can also be used to quantify the degree of soil compaction. However, interpreting bulk density as an indicator of compaction requires knowledge of soil texture — the percentage sand, silt, and clay — because texture influences the range of bulk densities considered normal or restrictive for a given soil. For example, a clay soil with a bulk density of 1.4 g/cm³ may feel hard and compact, especially when dry, yet a loose coarse sandy soil may have a bulk density of 1.7 g/cm³ without being compacted. This apparent paradox reflects the textural differences in porosity discussed in Section 2.3: clay soils have inherently higher total porosity than sandy soils, so the threshold bulk density at which compaction restricts root growth differs between textural classes. Table 3 provides rough guidelines for the minimum dry bulk density at which root growth is affected for various soil textures.

Table 3. Bulk density and root growth relationships based on soil texture.

Soil Texture	Ideal bulk density (g/cm³)	Marginal bulk density (g/cm³)	Root restricting bulk density (g/cm³)
Sands, loamy sands	<1.60	1.69	>1.80
Sandy loams, loams	<1.40	1.63	>1.80
Sandy clay loams, clay loams	<1.40	1.60	>1.75
Silts, silt loams	<1.40	1.60	>1.75
Silt loams, silty clay loams	<1.40	1.55	>1.65
Sandy clays, silty clays, clay loams	<1.10	1.49	>1.58
Clays (>45% clay)	<1.10	1.39	>1.47

Source: USDA-NRCS

[4]The soil was a Rains sandy loam 15-25 cm deep taken from a field in November that had been tilled six months before with a subsoiling chisel, tandem disk and field conditioner.

[5]In comparing results from different labs, always verify the analytical methods and rationale. North Carolina Department of Agriculture (NCDA) soil tests are conducted on a mass-per-unit-volume basis, reported as sufficiency index values for easy interpretation. Conversion to an area equivalent is based on a depth of 20 cm (7.9 inches) and a bulk density of 1 g/cm³. See end note "Conversion" for full rationale. Factors converting elemental nutrients to oxide (P, K) and ion (N) basis are also used to express fertilizer equivalents.

Compaction occurs when soil particles and aggregates are pressed together into a smaller volume. For example, if we filled a bucket with soil and packed it down so that it occupied one-half the volume, the soil's mass per unit volume must increase. This is the compaction procedure (Figure 4).

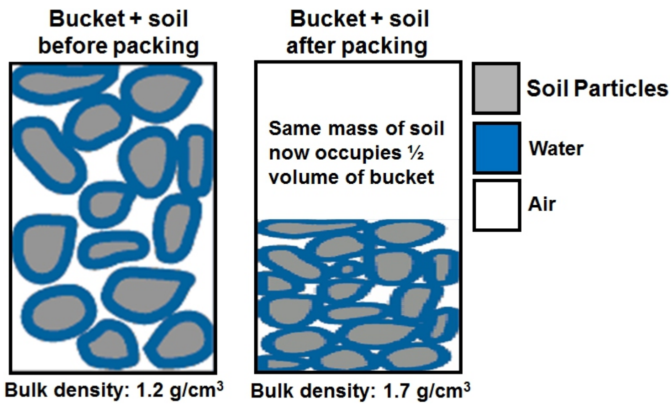


Figure 4 General relationship between packing density and volume for two levels of bulk density.

Soil compaction is conditioned on two principal agents:

- **Natural** Soil particle size distribution properties; pressure from root penetration; kinetic energy of rainfall and infiltrated water; overburden pressure; extent of waterlogging and freeze-thaw cycles.
- **Artificial** Field machinery pressure, number of passes, drive slippage; actions related to tillage and crop residue management; mistakes in crop rotation such as continuous cropping.

As compaction increases, the volume of air-filled porosity in the soil decreases, and its density, or mass per unit volume, increases.

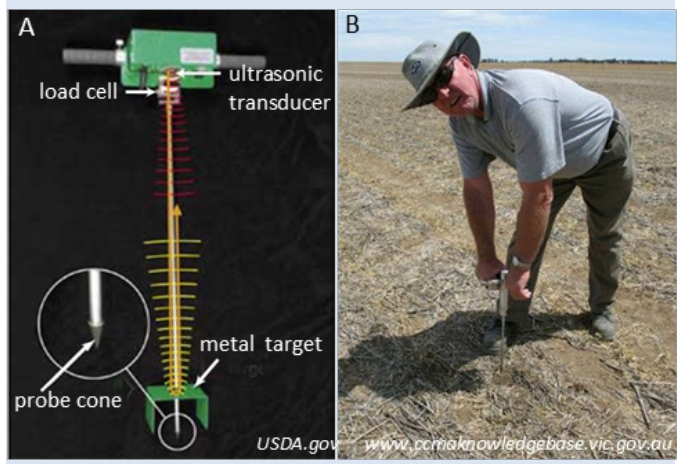
Harking back to our bucket of soil, it is evident that the packed soil occupies a smaller volume, yet the soil's mass in the bucket remains unchanged. The soil is simply denser than before; its response to packing was the expulsion of air and the rearrangement of solid particles into closer contact. In practical terms, compaction is the rearrangement of soil particles *under loading* that decreases the fraction of air-filled pores.

Soil compaction can be either good or bad. From an agricultural point of view, excessive compaction is bad because it reduces seedling emergence and vigor, root growth, infiltration and internal drainage, decreases aeration and increases the energy costs of tillage (Figure 5; see note 2.41).

Figure 5 Root deflection caused by soil compaction. Shallow rooting limits nutrient and water access by plants, so reducing yield. Escaped nutrients go to waste or worse, may become pollutants. The plants pictured here are "tillage" radish (*Raphanus sativus*), a strain of daikon radish employed to break up hard pans in soil. These specimens were evidently having trouble penetrating a 15-cm deep traffic pan with a measured bulk density of 1.82 g/cm^3. Tillage radish is not invincible! *Source: R.Walters, N.C. State Univesity.*

Note 2.41

Bulk density is measured to evaluate the level of soil compaction. However, plant roots do not respond directly to bulk density. The rate of root extension, and increase in root diameter, is governed by soil mechanical impedance which is conditioned on particle size, pore space geometry and water content. Generally, as bulk density increases and soil water decreases, mechanical impedance increases. Water in the soil acts as a lubricant, allowing plant roots to extend even through very small pore spaces. The exact change in mechanical impedance per unit change in bulk density and water content is soil dependent. Mechanical impedance is measured in the field with a cone penetrometer (Insets A and B).



Inset (A) shows the principal operating components of a typical hand-held penetrometer, and (B) shows the field operation. The metal target kicks a signal back to the transducer, allowing the penetrometer to calculate the probe's depth. Cone penetrometer readings may be expressed in kilograms-force per square centimeter (kg-f/cm^2: aka "cone index") or converted to pressure units, kilopascal (symbol: kPa).

Properly calibrated, cone penetrometers are useful for detecting zones of compaction or loosening in the subsoil.

On the plus side, some compaction is needed to ensure proper seed-soil contact during planting. Modern row crop planters are equipped with trailing press wheels for this purpose (Figure 6). Down pressure from the press wheels closes the planter disk opening, providing contact between the seed and surrounding soil particles. Compaction also increases the soil's mechanical strength, providing anchorage for a plant throughout its life cycle.

Figure 6 Planter press wheel ensures favorable seed-soil contact by firming the soil 360 degrees around the seed after it is placed in the slit opening. Target soil density of ~1.3 g/cm^3 surrounding the seed provides optimal movement of nutrients, water, and air through the soil. *Source: CaseIH.com*

Soil compaction in traffic lanes increases machine traction and reduces the horsepower needed for tillage. Compaction ensures the stability of earthworks formed from transported fill material placed beneath buildings, roads, bridges, power plants, etc. Unforeseen changes in soil volume beneath a structure can result in cracking, shifting, sagging, or in extreme cases, failure.

As the weight of modern farm machinery increases, problems related to soil compaction have become more urgent. The shift toward labor- and energy-saving conservation tillage farming means there is less soil disturbance through tillage manipulation. Less soil disturbance is generally viewed as beneficial for conservation, but unless machinery traffic is controlled, the benefits may be offset by compaction.

2.5 Particle Size Distribution

The compactibility of soil is strongly influenced by its three primary mineral components: sand, silt, and clay. The relative proportions of sand, silt, and clay define soil texture, which can be estimated in the laboratory by particle size analysis or assessed in the field by how the soil feels. A gritty soil has a high percentage of sand; a silky soil is high in silt; a soil that can be molded into a ribbon is high in clay. Particle size distribution may vary with depth and location within a field, which is an important consideration when evaluating soil compaction, since the bulk density thresholds associated with compaction differ among textural classes.

Figure 7 shows seven soil textures and their particle size distribution profiles. While these figures are considered 'typical', the sand fraction in a sandy loam soil may vary from ~50–80%. Conversely, soils with more than 20% clay are loamy in texture, and those with more than 45% clay are called clay soils. Even small amounts of clay disproportionately influence soil physical and chemical properties compared to the same amount of sand and silt.

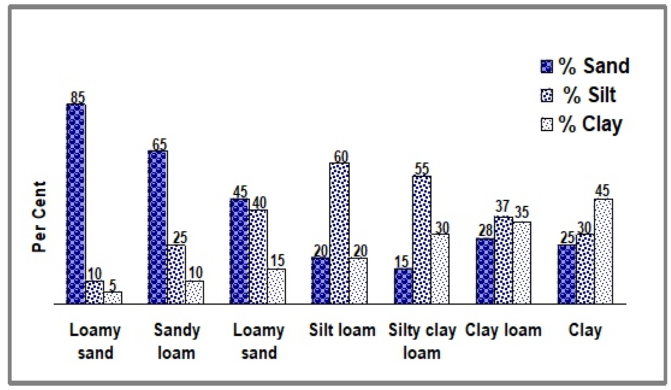


Figure 7 Percentage sand, silt, and clay for seven soil textural classes. *Source: redrawn and slightly modified from Siemens et al. John Deere Publ. (1993).*

The packing density of soil particles depends largely on particle shape and the proportions of different particle sizes present in the soil. Sand particles are typically spherical, though they range from well-rounded to angular depending on their mineralogy and transport history. The shape of clay particles is plate-like. Silt particles are intermediate in size between sand and clay and are often approximated as spherical for modeling purposes, though they are more irregular than idealized spheres and can consist of feldspars, micas, carbonates, and other minerals in addition to quartz. A soil containing a mixture of sand, silt, and clay can be packed together in a very dense configuration (Figure 8a) because the smaller silt and clay particles occupy the pore space between the larger sand particles. The extent of contact surfaces between individual particles is maximized, resulting in a very dense, mechanically stable mass with minimum pore space.

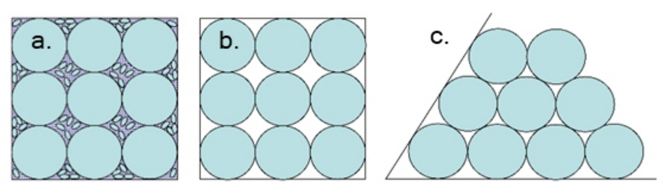


Figure 8 Packing density of soil particles: (a) mixture of sand, silt, and clay particles with capillary micropores; (b) uniform sand particles with non-capillary macropores; (c) the density of uniform sand is independent of packing arrangement.

Note, however, that maximum packing density does not occur at equal proportions of each particle fraction. The proportions required for smaller particles to optimally fill pore spaces between larger particles are governed by geometry. On the other hand, a sandy soil consisting of many grains of uniform size would be more difficult to compact because the geometry of the sand particles does not lend itself to tight packing. With no smaller particles present to occupy the voids between grains, many large pores remain between the sand grains regardless of arrangement (Figures 8b and 8c), limiting the degree of compaction achievable under moderate compactive effort.

Several important agricultural soils in the southern U.S. Coastal Plain have shallow, subsurface "pan" layers consisting of naturally compact, sandy loam and loamy sand "E" horizons (Figure 9). These light-colored, moderately well- to well-drained soils are typically low in clay and organic matter, with a diverse particle-size distribution in the sand fraction. What is the reason for such "genetic" hard pans?

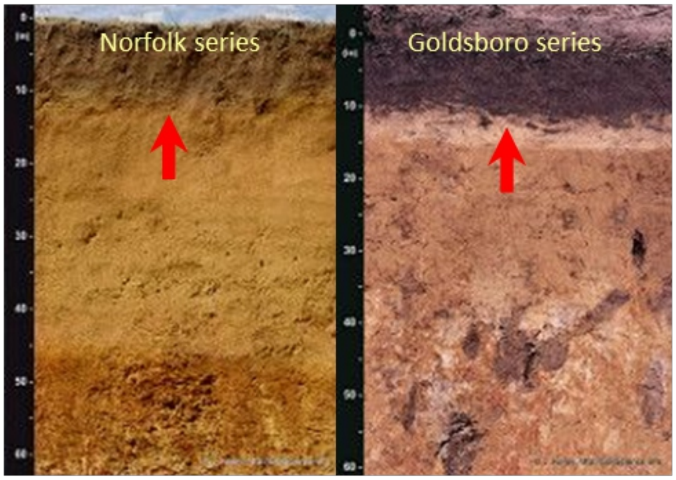


Figure 9 Two related southern U.S. Coastal Plain soil series with E horizons. The E horizon begins at the base of the darker-colored topsoil layer and extends several inches into the lighter-colored sandy soil beneath. Zones of compaction in the subsoil cannot be diagnosed solely from surface texture. *Image source: J. Kelly SoilScience.info*

Again, we must reach back to particle size distribution for understanding. Diversified particles in the sand fraction promote tight packing, creating an interlocking mass with very little pore space, high bulk density, and internal strength. Soils with genetic E horizons may require subsoiling periodically to maintain productivity, though its frequency and long-term efficacy depend on soil texture, ripper point design and working depth.

2.6 Soil Water-Density Relationships

Water in the soil has a lubricating effect, making it easier to press soil particles and aggregates together (compaction), or to pull them apart (tillage). It is easier to compact moist soil compared with dry soil. It is also easier to plow moist soil than dry soil. The optimum moisture content for compaction varies with soil. For a given particle-size distribution, the pressure required to compact soil decreases with increasing moisture content until an optimum value is reached. Beyond this optimum, additional water occupies pore space that cannot be expelled under the applied load, and the bulk density achievable for a given compactive effort declines. Figure 10 portrays the relationship between soil dry bulk density and water content.

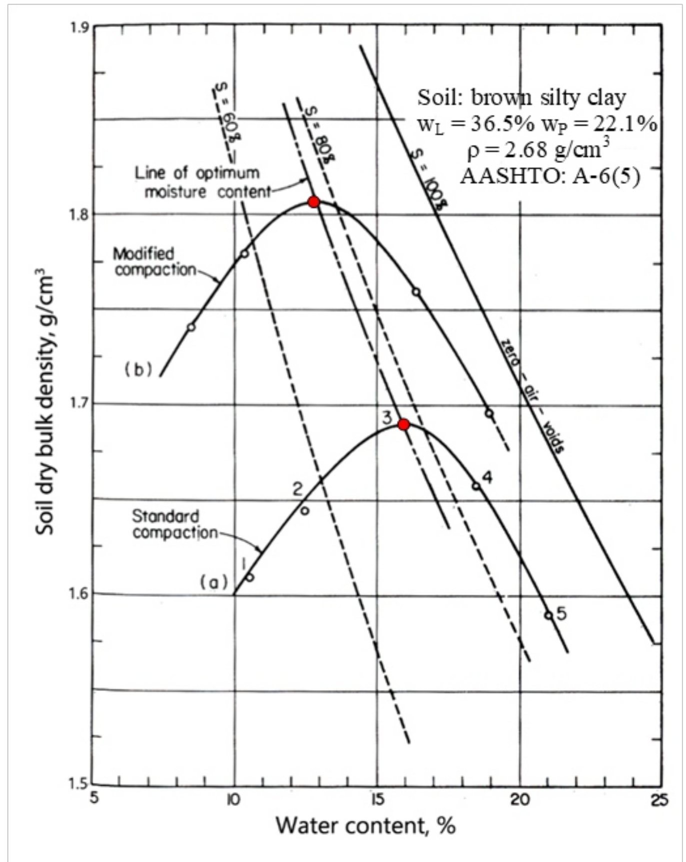


Figure 10 Relationship between soil dry bulk density and water content at two compaction levels. Red dots point to the maximum dry density. The figures wL and wP are the upper and lower plasticity limits ("Atterberg" limits), i.e. moisture content where the soil behaves as a moldable plastic. S=degree of saturation; ρ=particle density defined in Section 2.1. *Source: adapted from Bowles, 1984.*

The two compaction curves in Figure 10 are five-point Proctor density curves, named after R.R. Proctor, who first developed them for control specifications of dam construction earthworks for the Los Angeles Bureau of Waterworks and Supply in the early 1930s. The point of maximum density is reached at a critical optimum moisture content, which depends on the compaction pressure. As moisture content increases above the optimum for maximum density, additional water occupies pore space that cannot be expelled during compaction. Because water is effectively incompressible under normal compaction pressures, dry bulk density decreases beyond the optimum moisture content. Higher pressure ('Modified compaction') produces maximum density at lower moisture content compared with lower pressure ('Standard compaction'). In practical terms, this means it takes more effort to compact dry soil and, at all pressures, the greatest compaction occurs at a soil moisture content near the optimum.

As such, soil compaction susceptibility is greatest when moisture content is near or somewhat below the Proctor optimum, which for many agricultural soils approximates field capacity, though the precise relationship depends on soil texture and structure. Above field capacity, volumetric compaction is reduced because water-filled pore space cannot be compressed. However, wet soils remain highly vulnerable to structural damage through a different mechanism: shear failure. Unlike compaction, which reduces pore volume through particle rearrangement, rutting under wet conditions primarily results from shear failure, in which soil particles and aggregates are displaced laterally rather than pressed together (Figure 11). Both processes degrade soil structure through different mechanisms and require different management responses. Dry soil is also susceptible to compaction under loading, but it takes considerably more pressure than at higher moisture content.

Figure 11 Field rutting from machinery traffic. As soil is saturated above field capacity, it becomes less compressible. Red arrows point to "side walls," i.e., soil that has been sheared away from the center footprint. Annual tillage can smooth out field ruts. Rutting is more problematic under minimum and conservation tillage planting systems.

Core Intelligence

Conversion: As footnoted in Section 2.3, North Carolina Department of Agriculture (NCDA) soil tests are made on a mass per unit volume basis. Conversion to an area equivalent is based on a depth of 20 cm (7.9 inches) and a bulk density ("weight per unit volume") of 1 g/cm^3. Since the area of one hectare is 10,000 m^2 or 1 million dm^2, the volume of one hectare to a depth of 20 cm or 2 dm equals 2 million dm^3. Under this rationale, 1 mg/dm^3 = 1 ppm. To convert nutrients to *kilograms per hectare* (kg/ha, area basis), mg/dm^3 is multiplied by 2. To convert to pounds per acre (lb/acre) multiply kg/ha by a factor of 0.892.

Field capacity: Field capacity is the water content of a soil after the rate of downward drainage has become negligibly small following saturation. Following irrigation or rainfall, water moves downward under gravity; as larger, well-connected pores drain, hydraulic conductivity decreases sharply, and the remaining water is retained in smaller pores by capillary forces governed by adhesion, surface tension, and pore geometry. The matric potential at field capacity is conventionally approximated at −10 kPa for coarse-textured soils and −33 kPa for finer-textured soils. Most agricultural soils reach field capacity within 24–72 h after wetting, though this may be shorter in coarse sands or extended in fine-textured or layered profiles.

Soil compactibility*:* Soil compaction, soil compressibility, and soil compatibility are terms that are often used indistinctly in literature. Here, soil compaction is defined as a procedure that decreases the volume of air-filled pore space by applying an external load. Soil compactibility is the sensitivity of a soil to change in air-filled porosity per unit change in loading stress. Soil compressibility describes the sensitivity of a soil to changes in pore space per unit change in loading stress, irrespective if water or air is expelled from the pore space. Soil consolidation is a procedure wherein change in pore volume, or void ratio, is coupled with expulsion of excess pore water from the soil body. We do not consider soil consolidation in this technical note.

Further Reading

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