

PORE-SIZE DISTRIBUTION

showing that there was more than one type of binding agent in that soil.

Soils with high contents of organic matter have good moisture-retaining capacities, which has been attributed to the soil polysaccharide fraction and in particular to the highly viscous mucilaginous polysaccharides produced by soil microflora. The uronic acids in soil polysaccharide contribute to the cation exchange capacity of the soil. Similarly, in acidic soils, amino groups may contribute to the anion exchange capacity. The carbohydrate in soil seems to be the most important energy substrate for soil organisms.

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PORE-SIZE DISTRIBUTION

Porosity and pore-size distribution are important soil properties in both the disposition of water falling on the earth and in the growth of plants. Soils with large and stable pores absorb rainfall and permit it to percolate (see *Percolation*) downward rather than flow away over the surface. Maximum plant growth depends on suitable distribution of large, intermediate, and small pores. Predominance of small pores favors waterlogged conditions and poor aeration, whereas too many large pores make soils droughty.

The nature of pores in soils, especially those in uppermost layers, can affect air temperatures (see *Thermal Regions*). A dry and fluffy surface layer with many large pores is a good insulator, causing incoming solar energy to be reflected and reradiated rather than absorbed. The reflected solar energy heats the air, which in turn helps promote the turbulence enjoyed



FIGURE 1. Soil pore: Electron micrograph showing a 10^{-3} mm pore with amorphous silica acid coating (arrow). (From Uehara and Jones, in Cary and Evans, 1975)

by glider pilots on warm summer days. In contrast, a moist and dense surface layer with predominantly small pores is a good conductor of heat and encourages downward transfer of the incoming solar energy. Moreover, the heat stored during the warmest part of the day is released to the air during cool nights, enough at times to prevent frost when temperature of the air above the ground is near freezing.

A loam soil has 40–60% pore space between the mineral particles which, in their great variety of sizes and shapes, form the matrix around the pores (Figure 1). The pores are irregular in shape and range in average diameter from a millimeter to a few nanometers. Clay-textured or "heavy" soils have many pores less than 0.001 mm in diameter, whereas sandy or "light" soils have more pores larger than 0.001 mm.

Pore Sizes and Soil Water Storage

Water rises into a glass capillary tube according to the relation

$$h = \frac{2\tau}{\rho g r} \quad \text{or} \quad P = \frac{2\tau}{r} \quad (1)$$

where h is the height of the water rise; τ , the surface tension; ρ , the density of the liquid; g , acceleration due to gravity; r , the radius of the tube; and P , the air pressure above ambient

required in the capillary to prevent water from rising. This relation is used to measure pore-size distributions with the apparatus shown in Figure 2. The pores in the ceramic plate are so small that when they are filled with water, it cannot be displaced by the air pressure in the chamber. However, the water from larger pores in the soil can drain through the plate as the chamber pressure is increased. Measuring the incremental volumes of water outflow and using equation 1 leads to the release curves shown in Figure 3. These curves show both the water-storage capacities and the pore-size distributions of soils.

Pore Sizes and Soil Water Conductivity

Water flows through large soil pores rapidly and through small pores slowly. When the amount of water is not large enough to completely saturate the soil, the larger pores are not filled, and the water flows through the smaller ones, moving very slowly. A thin sheet of water standing on the surface during a rain is absorbed by a typical soil at a rate of about 1 cm/h, but 2 or 3 days after water disappears from the surface, flow rates in the soil decrease to less than 1 mm/day as the larger pores empty. Although infiltration rates of 1 cm/h are common, a change in the pore-size distribution at the surface may shift this value tenfold or more.

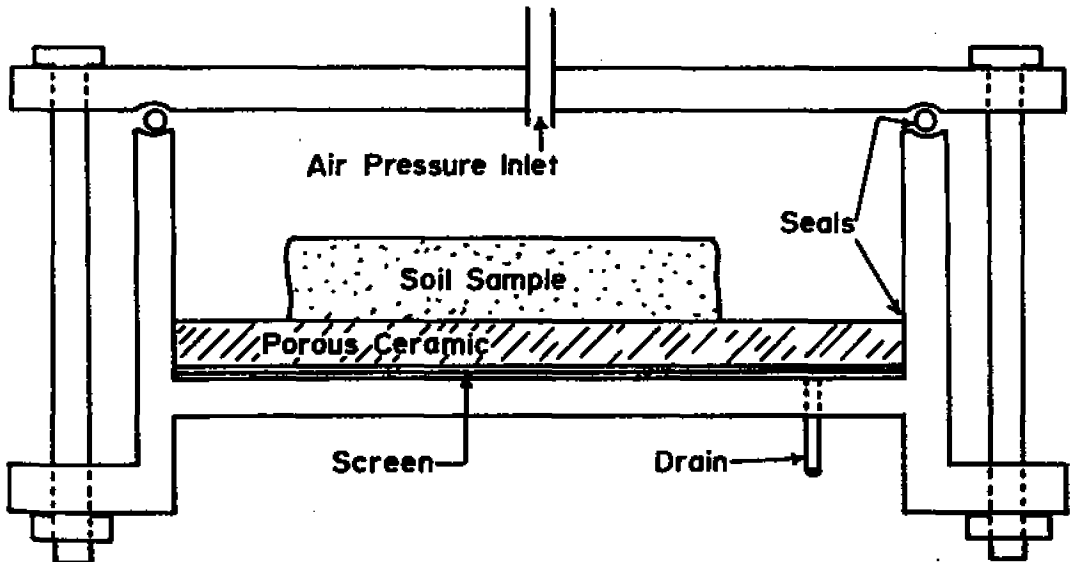


FIGURE 2. Diagram of a pressure chamber used to measure soil pore-size distributions and water-storage capacities.

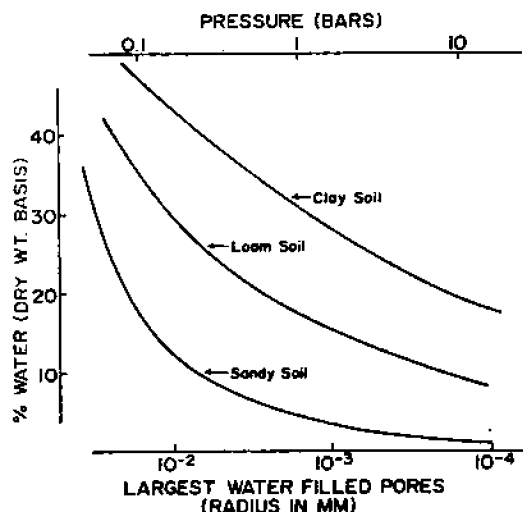


FIGURE 3. Typical relations between the soil water content, the largest water-filled pores, and the pressure in the apparatus shown in Figure 2.

The flow of water through soil is described by Darcy's equation:

$$J_w = -K \nabla \phi \tag{2}$$

where K is the hydraulic conductivity, and $\nabla \phi$ is the hydraulic gradient. Included in $\nabla \phi$ are both the gravitational and capillary pressure potentials. The capillary pressure P (as defined in equation 1) depends on the soil pore-size distribution.

Hydraulic conductivity is also affected by the soil pore-size distribution. Jackson (1972), summarizing the methods of Marshall, Millington, and Quirk, gives the relation

$$K_i = K_s \left(\frac{\epsilon_i}{\epsilon_s} \right)^n \frac{\sum_{j=1}^m [(2_j + 1 - 2_i)h_j^{-2}]}{\sum_{j=1}^m [(2_j - 1)h_j^{-2}]} \tag{3}$$

where K_s is the saturated conductivity; ϵ is the water-filled porosity; subscript i denotes the water content of interest; n is a constant; m is the number of increments of water content (or pore-size groups) over which the calculation is to be made; j is any particular increment; and h (as defined in equation 1) is related to pore sizes as the soil drains. Thus the conductivity can be calculated for any soil water content from the pore-size distribution as given in its moisture desorption curve (Figure 3) and a single measurement of its saturated hydraulic conductivity. There are also many techniques for measuring K directly. Figure 4 shows the relation between water-filled pores and conductivity. Conductivity decreases rapidly as the

soil becomes unsaturated, causing water to flow very slowly, even though hydraulic gradients may increase. This means that the many tiny soil pores store water like a blotter, only transmitting it downward when the soil is near saturation and the larger pores are filled (see *Water Fluxes*).

Porosity Effects on Air and Heat Transfer

Heat and gas transfer, which occur primarily by diffusion, are both affected by the total pore space, whereas pore-size distribution is relatively less important. Transport by diffusion is proportional to the porosity, but viscous flow is proportional to the square of the pore radius. The pore-size distribution primarily affects the distribution of water in soil, which in turn affects heat and gas transfer. Viscous flow of air, which also transports heat, occurs principally in the surface few centimeters of soil where pore-size distribution is important.

Soil thermal conductivities, used in the Fourier heat-flux equation, are most often calculated by the method of deVries (1963):

$$\lambda = \frac{x_0 \lambda_0 + k_1 x_1 \lambda_1}{x_0 + k_1 x_1} \tag{4}$$

where λ is thermal conductivity; the subscript 1 denotes the solid phase, and 0 the air or water phase that predominantly fills the pores; x the volume fraction; and k a factor to account for microscopic thermal discontinuities between phases. Thermal conductivity may range from 0.5 mcu/cm s°C for a dry porous soil to as much as 10 mcu/cm s°C for a wet, dense mineral soil.

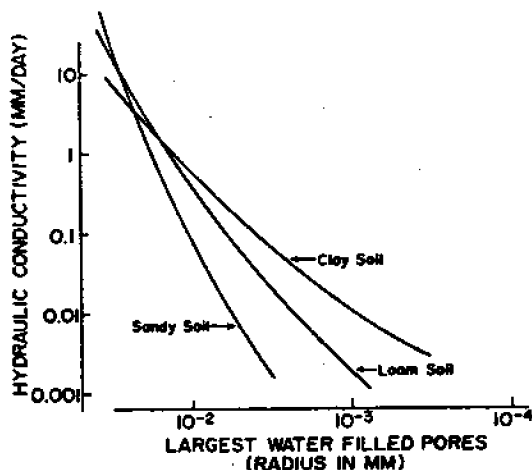


FIGURE 4. Typical relations between soil hydraulic conductivity and the largest water-filled pores.

Baver, Gardner, and Gardner (1972), in a review of the effects of porosity on soil gas diffusion, used the flux relation

$$J_g = D_0 S^n \nabla C \quad (5)$$

where D_0 is the diffusion constant of the gas in air; S is the air-filled soil pore space; n , a "constant" with values between 1 and 2; and ∇C , the concentration gradient.

Porosity Index and Plant Growth

A favorable soil pore-size distribution promotes vigorous plant growth. There must be enough large pores to allow rapid infiltration (q.v.) of surface water. These large pores must then drain quickly into smaller ones to allow aeration. Within a few hours after the surface is wetted, the soil must have at least 10% air-filled pores to dissipate carbon dioxide and provide oxygen for root respiration. There must be some pores with radii greater than 0.1 mm to encourage root elongation, as well as a large group with radii in the range of 0.01–0.001 mm to store soil water at an energy level easily available to plants.

A porosity index I (Cary and Hayden, 1973), may be calculated from the pore-size distribution to judge how well a soil meets these criteria, as

$$I = 0.67M + 0.75 \quad (6)$$

where M can be evaluated from three points on the soil moisture release curve (Figure 3). The index is in effect the dimensionless arithmetic mean change in water content as the pressure is raised from 0–1.5 bars in the apparatus shown in Figure 2. Most soils have an index less than 10, but when it falls below 2.5, the pore-size distribution becomes unfavorable for optimum plant growth.

Stability of Pore-size Distributions

Pore-size distribution is a dynamic soil property. Surface traffic, the weight of overburden, tillage (q.v.), root growth, soil fauna (q.v.), wetting and drying, and freezing and thawing continually rearrange the pores. The changes are greatest near the surface and in the pore-size groups with radii greater than 0.001 mm. Surface traffic decreases the proportion of large pores and increases the smaller ones. The single passage of a tractor tire will cause 80% of the total reduction in porosity that could occur with multiple trips. This reduction may be significant to a 30-cm depth, although most will be concentrated in the first 5–10 cm beneath the tire. Tillage, of course,

is often used to increase the number of large pores. Drying reduces the number of large pores, unless the soil is already severely compacted. In that case, wetting and drying and freezing and thawing tend to increase the proportion of pores with radii larger than 0.001 mm.

Soil particles are bound together by organic molecules, clay particles, the surface tension of adsorbed water films, and iron, aluminum, silicon, and lime cements. Consequently, increasing soil organic matter stabilizes soil particles and favors larger pore sizes. Increasing the amount of sodium, which disperses clay particles, and saturating soil with water, which eliminates surface tension forces, cause soil instability and a loss of the larger pore-size groups. Indeed, numerous factors influence soil pore-size distributions (Baver, Gardner, and Gardner, 1972; Gill and Vendenberg, 1967; and Cary and Evans, 1975), and many advances in soil management for maintenance of favorable porosity remain to be made.

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Cross-references: *Aeration, Respiration, Atmosphere; Bulk Density; Capillary Pressure; Conductivity, Hydraulic; Conductivity, Thermal; Diffusion Phenomena; Flow Theory; Soil Drainage; Soil Pores; Soil Structure; Thermal Regimes; Water Content and Retention; Water Fluxes; Water Movement.*

PORE SPACE, DRAINABLE

The *drainable pore space* of a soil or similar porous material is the difference between its volumetric water content at saturation or near saturation (below a water table or after a rain or irrigation) and its volumetric water content