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Porosity and Pore Size Distribution

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A soil's porosity and pore size distribution characterize its pore space, that portion of the soil's volume that is not occupied by or isolated by solid material. The basic character of the pore space affects and is affected by critical aspects of almost everything that occurs in the soil: the movement of water, air, and other fluids; the transport and the reaction of chemicals; and the residence of roots and other biota. By convention the definition of pore space excludes fluid pockets that are totally enclosed within solid material—vesicles or vugs, for example, that have no exchange with the pore space that has continuity to the boundaries of the medium. Thus we consider a single, contiguous pore space within the body of soil. In general, the pore space has fluid pathways that are tortuous, variably constricted, and usually highly connected. Figure 1 is an example of a two-dimensional cross section of soil pore space.

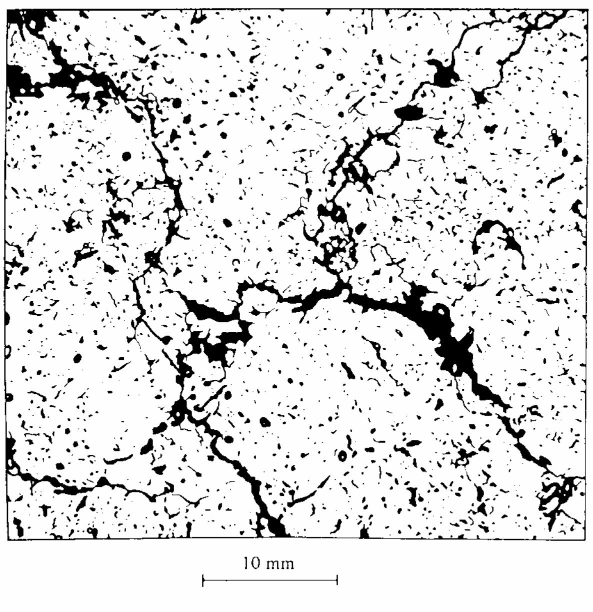

Figure 1. Cross section of a typical soil with pore space in black. This figure would lead to an underestimate of porosity because pores smaller than about 0.1 mm do not appear. (Adapted from Lafeber, 1965, Aust. J. Soil Res., v. 3, p. 143.)

The pore space is often considered in terms of individual pores--an artificial concept that enables quantifications of its essential character. Though many alternatives could serve as a basis for the definition of pores and their sizes, in soil science and hydrology these are best conceptualized, measured, and applied with respect to the fluids that occupy and move within the pore space.

Porosity

Porosity ϕ is the fraction of the total soil volume that is taken up by the pore space. Thus it is a single-value quantification of the amount of space available to fluid within a specific body of soil. Being simply a fraction of total volume, ϕ can range between 0 and 1, typically falling between 0.3 and 0.7 for soils. With the assumption that soil is a continuum, adopted here as in much of soil science literature, porosity can be considered a function of position.

Porosity in natural soils

The porosity of a soil depends on several factors, including (1) packing density, (2) the breadth of the particle size distribution (polydisperse vs. monodisperse), (3) the shape of particles, and (4) cementing. Mathematically considering an idealized soil of packed uniform spheres, ϕ must fall between 0.26 and 0.48, depending on the packing. Spheres randomly thrown together will have ϕ near the middle of this range, typically 0.30 to 0.35. A sand with grains nearly uniform in size (monodisperse) packs to about the same porosity as spheres. In a polydisperse sand, the fitting of small grains within the pores between large ones can reduce ϕ, conceivably below the 0.26 uniform-sphere minimum. Figure 2 illustrates this concept. The particular sort of arrangement required to reduce ϕ to 0.26 or less is highly improbable, however, so ϕ also typically falls within the 0.30-0.35 for polydisperse sands. Particles more irregular in shape tend to have larger gaps between their nontouching surfaces, thus forming media of greater porosity. In porous rock such as sandstone, cementation or welding of particles not only creates pores that are different in shape from those of particulate media, but also reduces the porosity as solid material takes up space that would otherwise be pore space. Porosity in such a case can easily be less than 0.3, even approaching 0. Cementing material can also have the opposite effect. In many soils, clay and organic substances cement particles together into aggregates. An individual aggregate might have a 0.35 porosity within it, but the medium as a whole has additional pore space in the form of gaps between aggregates, so that ϕ can be 0.5 or greater. Observed porosities can be as great as 0.8 to 0.9 in a peat (extremely high organic matter) soil.

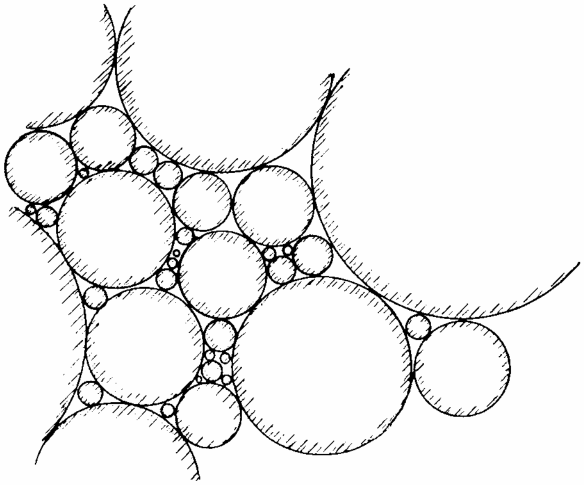

Figure 2. Dense packing of polydisperse spheres. (Adapted from Hillel, 1980, Fundamentals of soil physics, Academic Press, p. 97.)

Porosity is often conceptually partitioned into two components, most commonly called textural and structural porosity. The textural component is the value the porosity would have if the arrangement of the particles were random, as described above for granular material without cementing. That is, the textural porosity might be about 0.3 in a granular medium. The structural component represents nonrandom structural influences, including macropores and is arithmetically defined as the difference between the textural porosity and the total porosity.

The texture of the medium relates in a general way to the pore-size distribution, as large particles give rise to large pores between them, and therefore is a major influence on the soil water retention curve. Additionally, the structure of the medium, especially the pervasiveness of aggregation, shrinkage cracks, wormholes, etc. substantially influences water retention.

Measurement of porosity

The technology of thin sections or of tomographic imaging can produce a visualization of pore space and solid material in a cross-sectional plane, as in Figure 1. The summed area of pore space divided by total area gives the areal porosity over that plane. An analogous procedure can be followed along a line through the sample, to yield a linear porosity. If the medium is isotropic, either of these would numerically equal the volumetric porosity as defined above, which is more usually of interest.

The volume of water contained in a saturated sample of known volume can indicate porosity. The mass of saturated material less the oven-dry mass of the solids, divided by the density of water, gives the volume of water. This divided by the original sample volume gives porosity.

An analogous method is to determine the volume of gas in the pore space of a completely dry sample. Sampling and drying of the soil must be conducted so as not to compress the soil or otherwise alter its porosity. A pycnometer can measure the air volume in the pore space. A gas-tight chamber encloses the sample so that the internal gas-occupied volume can be perturbed by a known amount while the gas pressure is measured. This is typically done with a small piston attached by a tube connection. Boyle's law indicates the total gas volume from the change in pressure resulting from the volume change. This total gas volume minus the volume within the piston, connectors, gaps at the chamber walls, and any other space not occupied by soil, yields the total pore volume to be divided by the sample volume.

To avoid having to saturate with water or air, one can calculate porosity from measurements of particle density ρ_p and bulk density ρ_b. From the definitions of ρ_b as the solid mass per total volume of soil and ρ_p as the solid mass per solid volume, their ratio ρ_b / ρ_p is the complement of ϕ, so that

$$\phi = 1 - \rho_b / \rho_p. \quad (1)$$

Often the critical source of error is in the determination of total soil volume, which is harder to measure than the mass. This measurement can be based on the dimensions of a minimally disturbed sample in a regular geometric shape, usually a cylinder. Significant error can result from irregularities in the actual shape and from unavoidable compaction. Alternatively, the measured volume can be that of the excavation from which the soil sample originated. This can be done using measurements of a regular geometric shape, with the same problems as with measurements on an extracted sample. Additional methods, such as the balloon or sand-fill methods, have other sources of error.

Pores and Pore-size Distribution

The nature of a pore

Because soil does not contain discrete objects with obvious boundaries that could be called individual pores, the precise delineation of a pore unavoidably requires artificial, subjectively established distinctions. This contrasts with soil particles, which are easily defined, being discrete material objects with obvious boundaries. The arbitrary criterion required to partition pore space into individual pores is often not explicitly stated when pores or their sizes are discussed. Because of this inherent arbitrariness, some scientists argue that the concepts of pore and pore size should be avoided. Much valuable theory of the behavior of the soil-water-air system, however, has been built on these concepts, defined using widely, if not universally, accepted criteria.

A particularly useful conceptualization takes the pore space as a collection of channels through which fluid can flow. The effective width of such a channel varies along its length. Pore *bodies* are the relatively wide portions and pore *openings* are the relatively narrow portions that separate the pore bodies. Other anatomical metaphors are sometimes used, the wide part of a pore being the “belly” or “waist”, and the constrictive part being the “neck” or “throat”. In a medium dominated by textural pore space, like a sand, pore bodies are the intergranular spaces of dimensions typically slightly less than those of the adjacent particles. At another extreme, a wormhole, if it is essentially uniform in diameter along its length, might be considered a single pore. The boundaries of such a pore are of three types: (1) interface with solid, (2) constriction— a plane through the locally narrowest portion of pore space, or (3) interface with another pore (e.g. a crack or wormhole) or a hydraulically distinct region of space (e.g. the land surface).

This cellular, equivalent-capillary conceptualization of pores is especially relevant to hydraulic behavior, as has been recognized for more than 70 years. The initial application was to Haines jumps, illustrated in Figure 3, still considered the basic phenomena of capillary hysteresis. The pore openings, which control the matric pressure P at which pores empty, are smaller than the pore bodies, which control the P at which pores fill. As the medium dries and P decreases, water retreats gradually as the air-water interface becomes more curved. At the narrowest part of the pore opening, this interface can no longer increase curvature by gradual amounts, so it retreats suddenly to narrower channels elsewhere. An analogous phenomenon occurs during wetting, when the decreasing interface curvature cannot be supported by the radius of the pore at its maximum width. The volume that empties and fills in this way is essentially an individual pore. Not all pore space is subject to Haines jumps—water remains in crevices and in films (not seen in Figure 3) coating solid surfaces. Various models and theories treat this space in different ways. By the definition above it is part of a pore in addition to the volume affected by the Haines jump.

Pores can be classified according to their origin, function or other attributes. A textural/structural distinction is possible, analogous to porosity. Intergranular pores are the major portion of soil textural porosity, as discussed above. Intragranular or dead-end pores (if not entirely enclosed within solid) might empty or fill with water, but without contributing directly to fluid movement through the medium. Interaggregate pores, including shrink/swell cracks, are common types of macropores. Intraaggregate pores may be essentially equivalent to intergranular pores within an aggregate. Biogenic pores, for example the channels left by decayed roots and the tunnels made by burrowing animals, are another common type of macropores in soils.

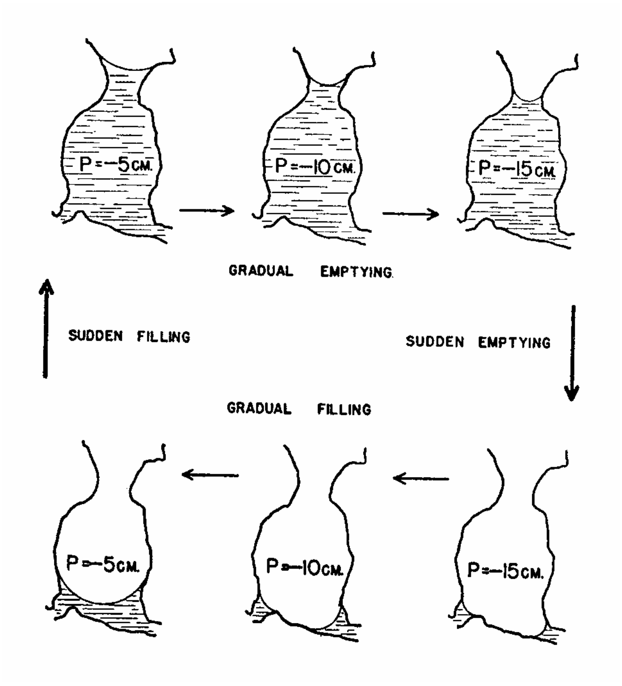

Figure 3. Dynamics of a Haines jump. (Adapted from Miller and Miller, 1956, J. App. Phys., v. 27, p. 324-332.)

Pore sizes are usually specified by an effective radius of the pore body or neck. The effective radius relates to the radius of curvature of the air-water interface at which Haines jumps occur. By capillarity this relates also to the matric pressures at which these occur, as discussed in the section below. Alternative indicators of size include the cross-sectional area or the volume of a pore, and the hydraulic radius, defined as the ratio of the cross-sectional area to circumference, or of pore volume to specific surface.

The pore-size distribution is the relative abundance of each pore size in a representative volume of soil. It can be represented with a function f(r), which has a value proportional to the combined volume of all pores whose effective radius is within an infinitesimal range centered on r. Figure 4 shows examples, all of which were derived from water retention data, as explained below. Like porosity, f(r) may be taken to comprise textural and structural components.

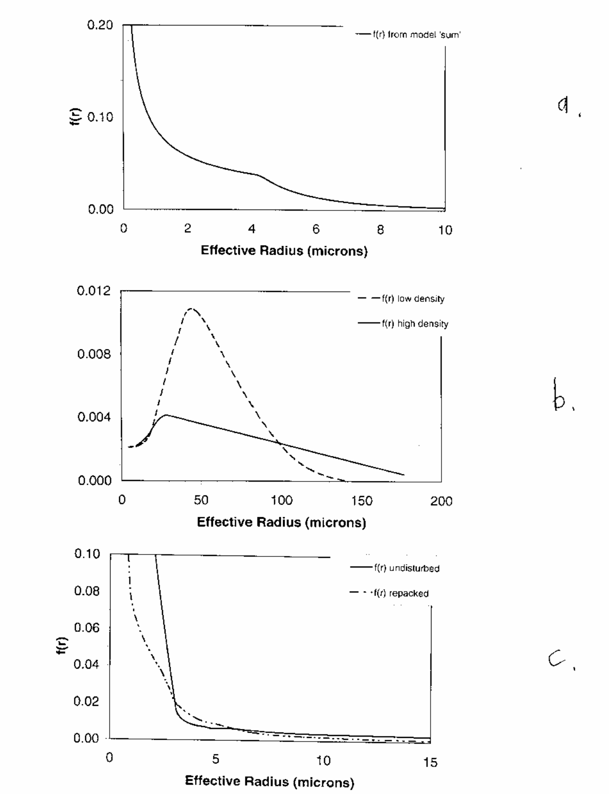

Figure 4. Pore size distributions based on measured water retention. (a) Loamy soil (Schofield, R.K., 1935, The pF of the water in soil, Transactions, 3rd International Congress of Soil Science: London, Murby & Co., p. 38-48). (b) Silty sand at two packing densities (Croney, D., and Coleman, J.D., 1954, Soil structure in relation to soil suction (pF): Journal of Soil Science, v. 5, p. 75-84). (c) Paleosol of sandy loam texture from 42 m depth, as a minimally disturbed core sample, and after disaggregation and repacking to the original density (Perkins, K.S., 2003, Measurement of sedimentary interbed hydraulic properties and their hydrologic influence near the Idaho Nuclear Technology and Engineering Center at the Idaho National Engineering and Environmental Laboratory: U.S. Geological Survey Water-Resources Investigations Report 03-4048, 18 p.).

Measurement

The most obvious and straightforward measurements of pore size are with geometric analysis of images of individual pores. This can be done using various types of microscopy on thin sections or other flat soil surfaces, or tomographs. Dimensions of pore bodies and necks can be measured manually or by computer analysis of digitized images. The lengths of segments of solid or pore along one-dimensional transects can serve similar purposes. As in the case of porosity, isotropy is required for assuming the equality of lineal, areal, and volumetric pore size distribution. For pore size, however, a more important problem is that when working with fewer than three dimensions, one doesn't know what *part* of the pore the selected slice intersects; because it does not in general go through the widest part, it underestimates the pore radius. Mathematical correction techniques are necessary to estimate unbiased pore body and opening sizes.

Three-dimensional analysis is possible with impregnation techniques. In these, the soil pore space is filled with a resin or other liquid that solidifies. After solidification, the medium is broken up and individual blobs of solid resin, actually casts of the pores, are analyzed as particles.

Image-based techniques can be prohibitively tedious because enough pores must be analyzed to give an adequate statistical representation. They can give a wealth of information, however, on related aspects such as pore shape and connectivity that is not obtainable otherwise.

More common than imaging methods are those based on effective capillary size. These use data derived from fluid behavior in an unsaturated medium, usually the emptying or filling of pores during soil drying or wetting (Figure 3). In other words, they use the water retention curve θ(P), where θ is the volumetric water content. Because large pores fill or empty at P near 0, a medium that has many large pores will have a retention curve that drops rapidly to low θ at high matric potentials. Conversely, one with very fine pores will retain much water even at highly negative P, thus having a retention curve with more gradual changes in slope. By capillary theory, the P at which a pore empties (or fills) corresponds to the pore opening (or body) size according to

$$r = \frac{-2\sigma\cos\alpha}{P} \quad (2)$$

where σ is the surface tension and α is the contact angle. This formula can convert a measured θ(P) into an equivalent θ(r) curve. This curve is actually a cumulative pore-size distribution; the water content on a drying θ(r) indicates the combined volume of all pores with opening radius less than r. Applying the fundamental theorem of calculus, the direct pore size distribution is simply the derivative:

$$f(r) = \frac{d\theta}{dr}. \quad (3)$$

Mercury porosimetry is analogous to the water-retention based method, but uses air as the wetting fluid, corresponding to the water, and mercury as the nonwetting fluid, corresponding to the air in the water-air system. Mercury is forced into the pores of dry soil incrementally, so that the relation between mercury content and mercury pressure can be recorded. Applying (2) with the appropriate values of surface tension and contact angle for mercury leads to a pore size distribution estimate as for the extraction of water from a water-air system.

The different measurement techniques do not give exactly the same results. Emptying and filling depends on more than capillarity – different methods are affected differently by unintended influences. Contact angles, for example, are dynamic, are not likely to be at handbook values in soil, and may deviate quite differently for water and for mercury. The swelling of clays can be a major influence on a water retention curve, but should have no effect with mercury. Imaging techniques are subject to some entirely different influences such as the chance and subjectivity involved in assessing pore bodies and openings. Figure 5 shows an example of substantially different results from different methods. The mercury porosimetry method indicates smaller pores than the photomicrographic method, in part because it gives a measure of pore opening rather than body sizes, and because it relies on dynamic accessibility of pores to the incoming fluid. In soil environmental applications, use of the water retention curve is most common. Data are more commonly available for this than for any other method. One can also expect a water-based method to give better results for a water-based application (e.g. hydraulic conductivity).

Representation

The basic equivalent-capillary representation of pore-size distribution is the function f(r) (Figure 4). Corresponding graphs of the

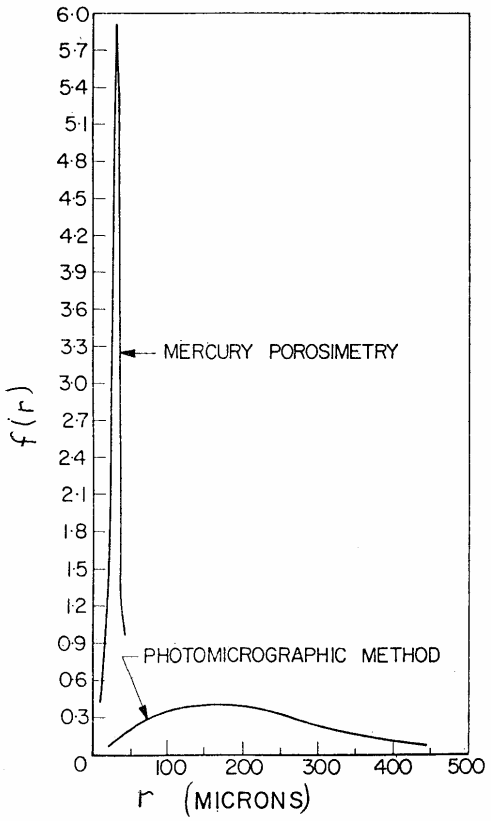

Figure 5. Pore size distributions obtained for an artificial medium using an imaging method and mercury intrusion (Dullien, F.A.L., and Batra, V.K., 1970, Determination of the structure of porous media: Industrial Engineering Chemistry, v. 62, no. 10, p. 25-53.).

cumulative size distribution can be used equivalently. Often a specific functional form or other representational model of pore-size distribution is useful. A normal or lognormal distribution can be fit to the data, for example. Structural features may give the soil a bimodal pore-size distribution, leading to several distinctive effects on water flow. Bimodal, trimodal, or other multimodal forms are possible, and can be represented by superpositions of the normal or lognormal forms. Self-similar or fractal forms of the distribution function are also used, giving a power-law form of the cumulative pore size distribution.

Typical features of a pore size distribution

With the equivalent-capillary concept taken as broadly as possible, the smallest possible pore is limited to the order of a few molecular diameters, about 1 nm. Fluid behavior in such pores is likely to be dominated by interaction with the solid material, and not necessarily describable by capillary laws or even standard thermodynamics. The smallest pores measured by water retention or mercury intrusion are typically 50 or 100 times larger than this, though with greater cost and effort, greater magnitudes of pressure and hence smaller measured r can be achieved. As the imputed pore size approaches zero, in reality the water is likely to be held in thin films that coat particle surfaces. The equivalent-capillary approach associates this water not with film thicknesses but with effective radii of hypothetical filled pores. Because r relates inversely to P, the whole dry portion of the retention curve is contained in a very small region near r=0, where the apparent number of pores becomes large. In the extreme case, when θ is held to remain artificially finite at an assumed residual water content as P goes to negative infinity, f(r) at r=0 is a delta function.

There is no real upper limit to pore size, though instability will cause the capillary hypothesis to break down for r of a few mm. The experimental limit is typically about 0.5 mm.

At either pore size extreme, where capillary phenomena lose dominance, Equation (2) no longer applies, though it still can give r values corresponding to P. These values may be useful for translating one property to another (discussed below) even though they are invalid in terms of the capillary analog.

At intermediate values of r, sometimes f(r) has a pronounced peak, as in Figures 4b and 5. Such a peak is assumed to exist in many applications and interpretations of pore size distributions, for example normal or lognormal representations, though it does not always exist in the measurable range of r. A peak in f(r) corresponds to an inflection point in θ(r). The inverse relation of r to P means that θ(P) is much more likely to have an inflection point than θ(r); if θ(r) has an inflection point, then θ(P) is mathematically required to have an inflection point, but the converse is not true. The common case of an retention curve that shows a wide range of slopes (with inflection point in the measured range, and normally a distinct initial-Haines-jump or air-entry effect) will have a defined peak in the pore-size distribution only if the middle portion of the retention curve has a slope that differs markedly from that of the two end segments. This is frequently true for monodisperse media and for repacked samples, in which pores larger than a certain r have been destroyed. For many soils, perhaps most, f(r) does not have a peak except at r=0 or so close to 0 that it cannot be measured. This is especially likely for media that are macroporous, polydisperse, or that show an air-entry effect at essentially P=0. Retention curves represented by a fractal or power-law model also are in this category. Lognormal models or other representations that have a defined peak can often still be applied but with the peak outside the measurable range. Some models (e.g. of hysteresis and hydraulic conductivity) may be unaffected by the lack of a measurable peak even though they were derived assuming the pore size distribution resembles some form of normal distribution.

Significance to soil and water behavior

Significance of porosity

One obvious significance of porosity is that it is an upper limit of volumetric water content. It is similarly essential to the definition of degree of saturation, θ divided by porosity. Another significance is that within the pore space, the complement to θ is gas content. That is, volumetric gas content equals the difference between θ and porosity.

The magnitude of porosity roughly indicates complexity of structure, being greater for greater complexity, as noted above for aggregated soil. The spatial variability of porosity is also important, greater variability correlating with greater heterogeneity of the soil. Significant spatial variability on a small scale also implies greater structural complexity, independent of the magnitude of porosity.

Significance of pore size distribution

A major importance of a soil's pore size distribution is that it relates to other soil properties in a complex and useful way. It indicates complexity of structure in far more detail than porosity alone. The spatial variation of pore size is an important characteristic of the medium. The pore size distribution of different parts of soil is the fundamental basis for the concept of aggregates, for example. By some definitions, pore size can permit essential distinctions between micropores and macropores (and mesopores, where that term is used for intermediate-size pores).

The relation of pore-size to particle-size distribution in a randomly structured medium is likely to be monotone: larger pores are associated with larger particles. The nonrandom structure of most soils adds complexity to the relation between pore size and particle size. Large pores can be associated not only with large particles, but also with smaller particles such as clays that promote aggregation and hence the existence of large interaggregate pores.

With more intricate conceptualizations, subdivisions of a pore size distribution can represent more detailed properties of soils, for example the hysteresis of soil water retention. Figure 6 shows an example in which one graph represents pore-opening and pore-body size distributions. Portions of the f(r) area are shaded differently, based on criteria of how the pore body radius (r_w in this diagram) relates to the pore opening radius (r_d in this diagram) and to r_s (defined on the abscissa). The added information of what fraction of pores with a given opening size have a particular range of body sizes is taken to represent the essential quantitative basis of soil water hysteresis.

Pore dynamics

Natural and artificial soil processes create and destroy pores, and induce changes in their size and other attributes. Table 1 lists some of the common effects on pore size distribution caused by routine processes. In general, if the soil is compacted by uniform stress, such as a weight imposed at the land surface, it normally loses large pores and gains small pores. Figure

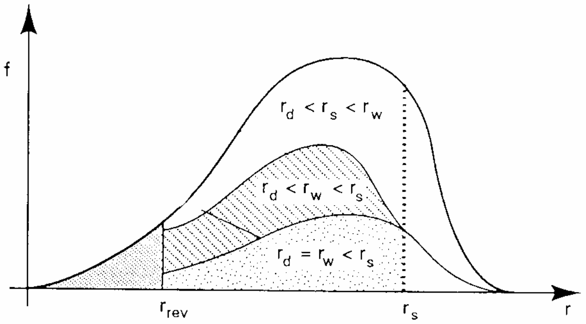

Figure 6. Hypothetical pore size distribution with subregions distinguished on the basis of pore body and opening radii and drying/wetting history (Adapted from Nimmo, 1992, Soil Sci. Soc. Am. J., v. 56, 1723-1730).

Table 1. Possible effects of routine soil processes on pore size distribution.

Shrinkage
• can enlarge macropores
• can create new macropores
• (within an aggregate) can cause intraaggregate pores to decrease in size, or to increase in size if clay particles are shrinking
Swelling
• can decrease the size of macropores
• can close macropores
• (within an aggregate) can cause intraaggregate pores to increase in size, or to decrease in size if clay particles are expanding
Mechanical compression
• can decrease the size of macropores
• can close macropores
• can break up aggregates, reducing the number of intraaggregate pores and thereby reducing the fraction of the pore space represented by the smallest pores
Disturbance from digging or plowing
• can destroy existing macropores
• can create interclod macropores
• can break up aggregates, reducing the number of intraaggregate pores and therefore also the fraction of the pore space represented by the smallest pores
Biological activity
• can create new macropores
• can enlarge macropores, as by ongoing traffic of ants or burrowing mammals
• can decrease the size of macropores, for example if they are affected by compression resulting from the expansion of a nearby root
• can increase aggregation, promoting the creation of interaggregate macropores, and possibly to smaller intergranular pores within aggregates
• can constrict or obstruct pores, for example by growth of microorganisms
Chemical activity
• can constrict or obstruct pores by formation of precipitates
• can enlarge pores by dissolution of precipitates
• can increase or decrease interparticle cohesion, with complex effects on pore size and structure

4b illustrates results of this type. Disturbance from irregular stresses, as during digging or repacking, has a variety of effects on pore size, often with the net effect of a decrease in the number of large pores and an increase in the number of small pores, as illustrated in Figure 4c. Several types of processes can create pores. Though small intergranular pores are seldom closed completely, some processes can close macropores, in effect destroying them.

Applications to soil transport properties

Because pores are fluid conduits, their size distribution is useful for predicting hydraulic conductivity K, as well as for water retention as described above. Gas and other types of fluid transport can be treated, though water flow is the most common application.

By analogy to laminar flow in tubes as quantified by Poiseuille's law, the conductance of a single pore can be inferred to be proportional to the fourth power of its effective radius. This makes its hydraulic conductivity proportional to the square of its effective radius. An estimated f(r) distribution indicates the relative abundance of each conduit size, thus providing the information needed to predict K.

A K prediction based on the capillary hypothesis assumes that the pores that are filled at a given θ have an effective radius smaller than a threshold determined from θ using the water retention relation. The portion of f(r) representing the filled pores is relevant to unsaturated K. The simplest possibility is to integrate f(r) weighted by r^2 over the domain represented by filled pores, giving a number proportional to K for the corresponding θ. Multiplying this integration by a separately determined matching factor gives the actual predicted K.

Many specific models in the published literature are based on these ideas. They differ in how they treat such matters as pore length, connectedness, tortuosity, and the distinction between pore opening and body dimensions. Popular models include those of Mualem and Burdine, which have been analytically combined with widely used empirical formulas for retention curves.

The pore size distribution affects solute convection similarly to hydraulic conductivity. Additionally it affects solute dispersion, which is expected to be greater for a broader pore-size distribution. It affects the sorption of solutes in a complex way. The smaller pores are associated with longer residence times and greater relative surface area, but most solutes may go quickly through the large pores with minimal opportunity to react. Interchange between fast-transporting and slow-transporting portions of the pore space is a vital and much-investigated aspect of solute transport in soils. Sometimes the terms “mobile” and “immobile” are used in this context, but of course the distinction is not as sharp as these terms imply.

For particle transport, many aspects are essentially the same as for K and solute transport. Additionally, the phenomenon of straining depends critically on the proportion of pores smaller than a given particle size. This is the dominant factor in some particle-transport applications.

Conclusions

The characterization of pore space is a vital and fruitful aspect of soil investigation. Liquid, solid and gas constituents of the soil govern the form and development of pores, whose character in turn profoundly influences the nature and behavior of the soil.

Soil porosity is fairly well standardized in definition and measurement techniques. Pore size, however, is not obvious how to define, much less to measure. Yet it is central to topics like macropores, aggregation, fractures, soil matrix, and solute mobility. Pore size plays a key role in various proposed means of quantifying soil structure. It also has a major practical role in the prediction of hydraulic properties. New pore size concepts, measurement techniques, and relations to transport phenomena are likely to remain a major emphasis in the study of soil.

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