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## HARDENING

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Hardening or induration of a soil takes place by the loss of void space by compaction or filling with fine materials.

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## HARDPAN

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A compacted, impermeable layer of soil at or near the surface.

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## HARDPAN SOILS: MANAGEMENT\*

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### Synonyms

Hard-layer soils; Management of hard-layered soils

### Definition

*Hardpan soil.* A soil that has a layer whose physical characteristics limit root penetration and restrict water movement.

*Penetration resistance.* The penetration resistance (or soil strength) is usually measured as the force exerted on

a standardized implement (penetrometer) as it is pushed into the soil divided by the cross-sectional area of its tip.

### Introduction

Hardpans, hard layers, or compacted horizons, either surface or subsurface, are widespread problems that limit crop production. Hard layers can be caused by traffic or soil genetic properties that result in horizons with high density or cemented soil particles (Hamza and Anderson, 2005); these horizons have elevated penetration resistances that limit root growth and reduce water and airflow. Limited root growth leads to limited crop water and nutrient uptake. Reduced water flow prevents rainfall or irrigation water from filtering into the soil profile where it can be stored for plant growth. Reduced airflow limits oxygen and carbon dioxide exchange with the atmosphere; exchange is needed for plant and microorganism respiration. These limitations reduce crop productivity.

Improving the hard layer consists of reducing its hardness or penetration resistance. When we reduce the layer's hardness, we assume that it and/or the layers below it have properties conducive to plant growth. As the hard layer softens, water and air are able to move into and/or through it and into the layers below, improving conditions for root growth and with its productivity. There are several ways to improve hard layers; the most common is tillage; but other solutions exist in the forms of water/crop management and soil amendments.

### Tillage

Tillage has been and is the common method used to remediate hard-layer problems; it physically breaks up hard layers. Tillage by hand involves digging with a spade, broad fork, or U-fork. In large-scale mechanical agriculture, tillage involves using a tractor to pull any of a number of tines or shanks through the soil. In the mechanical method, shallow hard layers (<5 cm) can be

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broken up with tines or cultivators that disrupt the surface soil. Deeper hard layers (>15 cm) can be broken up with shanks. Shanks are sized or adjusted so they are pulled through the soil at the depth of the hard layer shattering it and decreasing its resistance to root growth. Different shank designs that are manufactured by various tillage companies produce different results or work with different efficiencies depending on the type of hard layer and the type of soil. Consider the example seen in Figure 1 where the hard layer was located in loamy sand between 20-cm and 40-cm depths. Tillage in this example was performed with an older 5-cm thick shank that produced wider zones of disruption and used more energy than narrower shanks. Also seen in Figure 1, the process that reduced soil penetration resistance under the row increased it under the trafficked mid row because of the tractor weight.

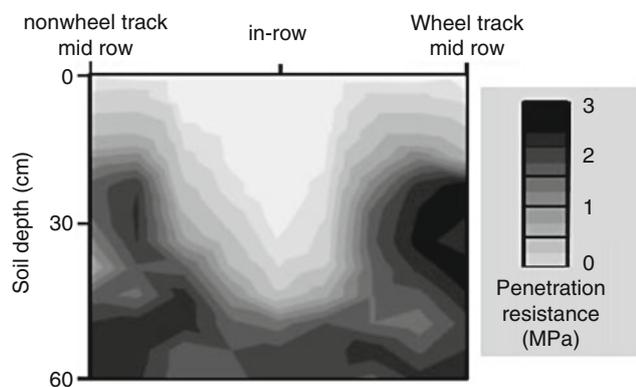
To meet conservation goals, deep tillage such as that shown in Figure 1 can be performed in such a way that it does not invert soil; equipment companies have developed shanks that break up soil with minimal surface disruption. Non-inversion tillage leaves most crop residue on the soil surface protecting it from erosion, surface crusting/compaction, and excessive evaporation (Raper, 2007). Though early studies with non-inversion and reduced tillage demonstrated little or no yield advantage, improvements in planters, residue management, and soil/crop management practices increased the success of conservation systems by optimizing factors that affected seed germination and vigor.

The problem with tillage is that the reduction of penetration resistance is temporary. For some soils, temporary means a few to several years. For others, it can mean only a few months (Raper et al., 2005a). Most often it is effective for only months. In either case, over time, soil reconsolidates leading to reduced water/airflow, reduced root growth, and lower crop yields (Håkansson and Lipiec, 2000). Even if the reconsolidated soil's penetration resistance is not as high as it was originally, it can be high enough to limit growth. As a result, tillage has to be

performed repeatedly at prescribed frequencies, often seasonally or annually. Frequent tillage can be expensive because it often requires large tractors (14–20 kg weight per shank), 20–40 min ha<sup>-1</sup> of labor, and 20–25 L ha<sup>-1</sup> of fuel. Eventually, the producer has to make the decision whether or not to till based on the value of increased yield by tillage vs. the cost of tillage (Bolliger et al., 2006).

In an effort to save time, fuel, and production costs, deep tillage studies have included soil disruption on a multiple-year rotation. In many cases, not tilling every year reduces yield to levels that may (or may not) be acceptable given the increase in fuel costs. Additionally, annual deep tillage may not be needed for some crops, such as cotton, to maintain yields. Deep tilling every 2–3 years may be just as effective as deep tilling annually (Busscher et al., 2010). This will depend on the crop and variety grown, amount of re-compaction, and other crop management techniques such as row width and traffic/compaction patterns.

Another effort to save fuel and production costs involves varying the tillage depth. Deep tillage is often performed with implements set to a fixed depth. But depth to the compacted layer varies throughout a field. What depth should the implement have? On the one hand, if tillage depth is based on the deeper zones of the compacted layer, the implement disrupts too much soil where the compacted layer is shallow; this wastes fuel. On the other hand, if tillage depth is based on the shallower zones, the implement will not disrupt the whole compacted layer, leaving hard zones that limit root growth. Technologies are now available that allow tillage to vary with the depth of the compacted layer; this can be accomplished by mapping the hard layer of a field or placing sensors on the shanks. Shanks are then raised and lowered as needed. This action can save energy without sacrificing crop yields. Research has shown that this “site-specific tillage” produced yields equivalent to those of uniform deep tillage while reducing tractor draft forces, drawbar power, and fuel used (Raper et al., 2005b).



**Hardpan Soils: Management, Figure 1** Soil penetration resistance for a loamy sand that has a hard layer at 20- to 40-cm depths. The soil was tilled to a depth of about 45 cm with shank that did not invert the soil.

## Other solutions

**Soil Organic Matter:** For the past few decades, soil scientists and producers have been trying to increase organic matter levels in soils (Carter, 2002). This improves fertility, decreases strength in hard layers (especially those close to the surface), and increases yield (Soane, 1990). But with the increase in fuel prices comes the need for organic matter/residue in the form of cellulose. The same organic matter that scientists and producers were trying to increase in soils may be removed to produce ethanol. Both increased organic matter and removed cellulose might be attainable; but only after some research. Research on organic matter removal had started during the 1970s fuel crisis; but because the crisis did not continue, the research priority decreased as funding ceased. Results from the 1970s showed that some residue could be removed provided that nutrients were replaced with fertilizers. The problem with this finding is that fertilizer

production requires large amounts of energy. The previously unfinished research has resumed asking questions about the sustainability, economic efficiency, energy efficiency of residue removal, and the effect of the removal on soil properties such as penetration resistance.

To ameliorate hard layers, additions of organic matter need not come from crop residues. Another way to add it, especially to subsurface hard layers, is through root growth (Yunusa and Newton, 2003). In this method, cover crops are grown between growing seasons aimed at penetrating the hard layers with their roots. Cover crop roots are able to penetrate soil where production crops cannot either because conditions between growing seasons are different, for example, cold and wet, or because the cover crop has hearty roots but it is not an economic crop. For example, large rooted crops such as radish are grown to add both large holes and large amounts of organic matter to the hard layer. In another method, rye cover crop roots penetrate compacted layers softened because they are wet in winter; rye roots leave holes behind for summer row crop roots to follow. Success of these methods depends on whether the roots can grow deep enough to affect the hard layer and whether or not the holes left by the roots collapse.

*Other management:* Another way to soften hard layers that met with some success was to irrigate the soil with drip tubes buried just above the hard layer. In this method, irrigation water keeps the hard layer soft while supplying the crop's needs. However, tubes buried above a hard layer require careful management to avoid overwatering or underwatering. Overwatering prevents roots from penetrating a flooded layer while underwatering does not loosen the hard layer enough. It is likely that both types of irrigation will occur simultaneously between and at the buried tubes or between and at irrigation ports or emitters along the tubes. Water management needs to find a proper irrigation schedule that can satisfy all needs for each soil. Because water is not at the soil surface, this type of irrigation reduces evaporation saving water but a relatively dry surface can reduce germination and stand establishment during years with early dry seasons.

A long-term solution that reduces both compaction and energy demand is to add amendments to soil. The amendment chosen will have to reduce compaction much like the organic matter does and it should be effective for several years because it will be expensive to incorporate it into the soil. Potential amendments include polyacrylamide (PAM) and biochar. PAM was tried in the 1950s. Older formulations were used to stabilize aggregates in the surface 30-cm to 40-cm depths. Hundreds of kilograms of PAM per hectare were needed, limiting PAM-use to high value crops and nurseries. Since the 1950s, polymer formulations and purity have improved, making them more effective at lower concentrations. In the 1990s, environmentally safe PAM was found to be an effective erosion-preventing and infiltration-enhancing polymer when applied at 1–10 mg L<sup>-1</sup> in furrow irrigation water. This can affect the surfaces of irrigated soils; but if the hard layer is deep

in the profile, the amount of PAM and its mixing into the soil will cost several hundred euros per hectare. Given the high cost of fuel, this cost might be feasible if the PAM could last multiple years. Current estimates have the PAM breaking down at a rate of 10% per year.

Another amendment that has attracted attention in the past few years is biochar. Biochar captured the attention of the agricultural community as a result of archeological/agricultural findings of charcoal-amended soils in the Amazon and other historically old areas. Charcoal- or biochar-amended soils were found to have supported larger populations 500–1,000 years ago than previously estimated and today they are still more productive than expected. If biochar can be effective over time and if it improves productivity, it could be economically feasible to use it as a long-term soil amendment to eliminate or reduce hard-layer tillage (Busscher et al., 2010). More research needs to be performed before making a final decision; but preliminary results are favorable. Biochars vary based on their source material and production technique. Current work is underway to match biochar properties to the needs of the soil and its hard layer; then their effectiveness needs to be assessed.

### Effects on individuals

Whether or not you work in tillage management or agrophysics, they affect you because of their impact on food, fiber, and energy production. As populations increase and as we make more demands on our resources, we will require tillage management and other areas of agriculture to produce more food for more people with a limited and dwindling soil base (Small, 2009). We can all become involved by being educated and active in conservation efforts to improve the lot of our soils, our environment, and our fellow men.

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### Cross-references

- Compaction of Soil  
 Conditioners, Effect on Soil Physical Properties  
 Hardsetting Soils: Physical Properties  
 Infiltration in Soils  
 Layered Soils, Water and Solute Transport  
 Root Responses to Soil Physical Limitations  
 Soil Surface Sealing and Crusting  
 Soil Penetrometers and Penetrability  
 Subsoil Compaction

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## HARDSETTING SOILS: PHYSICAL PROPERTIES

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### Synonyms

Cohesive soils or soils with a cohesive character

### Definition

*Hardsetting soils or soils exhibiting hardsetting behavior.* Soils that have horizons that, when dried, harden significantly, constituting a mass without structure (apedal); this soil till is more difficult or even impossible. This impediment can be avoided by humidifying the soils (Mullins et al., 1990; Mullins, 1999).

*Cohesive soils or soils with cohesive character.* Soils with dense pedogenic subsurface horizons, which are very resistant to penetration of the knife or hammer and are very hard to extremely hard when dry, becoming friable or firm when moist. In natural conditions, they have

a weak structural organization, generally appearing solid or with some tendency to form blocks (Fabiola et al., 2003; Lima et al., 2006).

### Introduction

Hardsetting is a phenomenon that occurs in many soils around the world in arid tropical, semiarid, and Mediterranean regions (Mullins, 1999) and covers more than 110 million ha of areas of agricultural exploitation. The term hardsetting was introduced by Northcote (1960) in binary textured soils of Western Australia, and was subsequently recognized in Africa, Asia, and South America (Mullins et al., 1987; Mullins et al., 1990; Chartres et al., 1990; Fabiola et al., 2003; Lima et al., 2006).

The Australians were pioneers in identifying and mapping hardsetting soils, as well as in incorporating these characteristics into a soil taxonomic classification system (Harper and Gilkes, 1994; Isbell, 1996). Nevertheless, the ambiguous nature of hardsetting behavior has limited the use of the term in other classification systems outside Australia and Brazil (Harper and Gilkes, 1994).

Many agricultural problems are associated with hardsetting soils, including a more restricted period for soil tillage and an increase in physical impediments to adequate root development (Mullins et al., 1987; Mullins et al., 1990). Hardsetting is normally associated with processes of soil degradation such as erosion, compaction, crusting, and acidification of the soil (Mullins, 1997). In these soils, the agricultural production is frequently frustrating due to low production and a high cost/benefit ratio.

### Characteristics of hardsetting soils

Hardsetting soils present a pedogenetic densification in the surface horizons (A and AB) and the subsurface horizons (BA, B, E, EB, BE) (Mullins et al., 1990; Chartres et al., 1990; Fabiola et al., 2003). When dry, they present a lack of visible structural organization (they are massive), elevated resistance to penetration by a knife or auger, and a hard to very hard (at times extremely hard) consistency. The humid soil consistency varies from friable to firm, and a dry sample, when immersed in water, disintegrates rapidly (Mullins, 1997).

Hardsetting characteristics normally occur in deep soils, with a loamy-sandy-clay texture, clay-like or very clay-like, in a plain to gently undulating relief. Hardsetting horizons possess soil bulk density higher than the underlying horizons and tensile strength values  $\geq 0.09$  MPa (Fabiola et al., 2003). From a chemical point of view, they present a low base saturation ( $V < 50\%$ ), organic material content  $< 2.0\%$ ,  $Fe_2O_3$  content (by  $H_2SO_4$ )  $< 8$  g  $kg^{-1}$ , and an illitic or kaolinitic mineralogy (Mullins et al., 1990; Giarola et al., 2001).

It is important to distinguish between hardsetting and compacted soils. Soil compaction results from repeated or long-term movement of agricultural machinery and stock compacting the soil profile when it is moist, often remaining hard when wet. In many cases, this compaction

layer occurs at depth. Hardsetting affects the A1 horizon (Mullins et al., 1987), but the soil softens when moist.

The hardsetting horizons should not be confused with fragipan, which also presents high levels of cohesion, but presents diverse pedogenetics (chemical grouting), occurring at greater depths; fragipan has different implications in relation to soil management (Chartres et al., 1990).

### Hardsetting processes

The hardsetting character can be associated with the following processes: (1) the precipitation of soluble salts in the contact zone between aggregates and/or soil particles (Mullins and Panayiotopoulos, 1984; Mullins et al., 1987; Mullins, 1999); (2) the dispersion of soil clay, associated or not with the presence of sodium (Mullins, 1999); (3) natural bulk density increases of the soil particles, which increase the effective stress and the water matrix potential in the soil as the soil dries (Fabiola et al., 2003) (Figure 1).

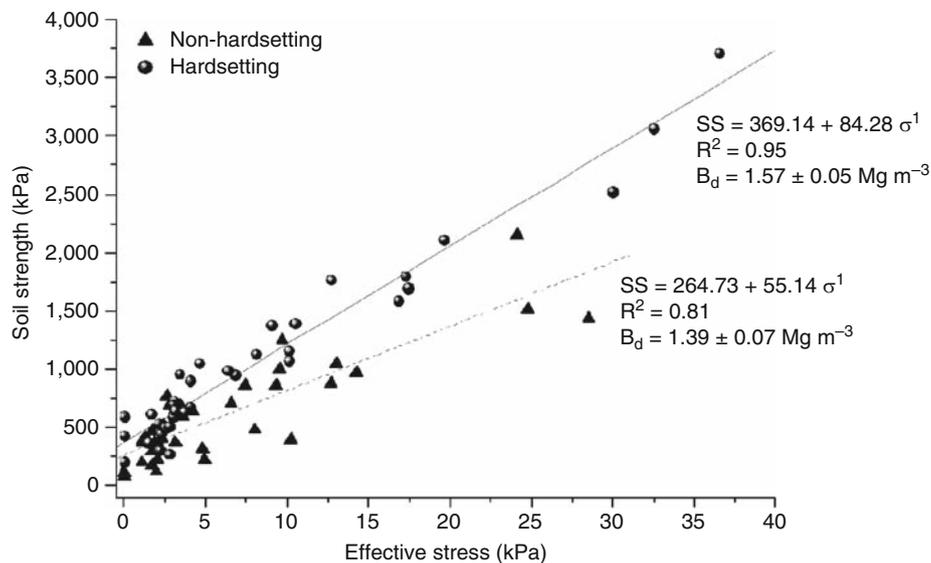
Other types of hardsetting soils have been distinguished in northern Cameroon (Lamotte, et al., 1997): (1) Soils with very hard sandy layers that usually occur under a more or less softer sandy layer. Some indications suggest that these properties could result from the gradual clogging of the pores between sand grains due to newly formed clay. Such a process may be favored by the succession of drying and wetting periods. (2) Soils with a very hard clay layer that are thought to be derived from Vertisols degraded by cultivation. These very hard layers, sandy or clayey, are not necessarily associated with a high sodium content but rather with a low iron content, as reflected by their pale color.

### Physical behavior of hardsetting soils

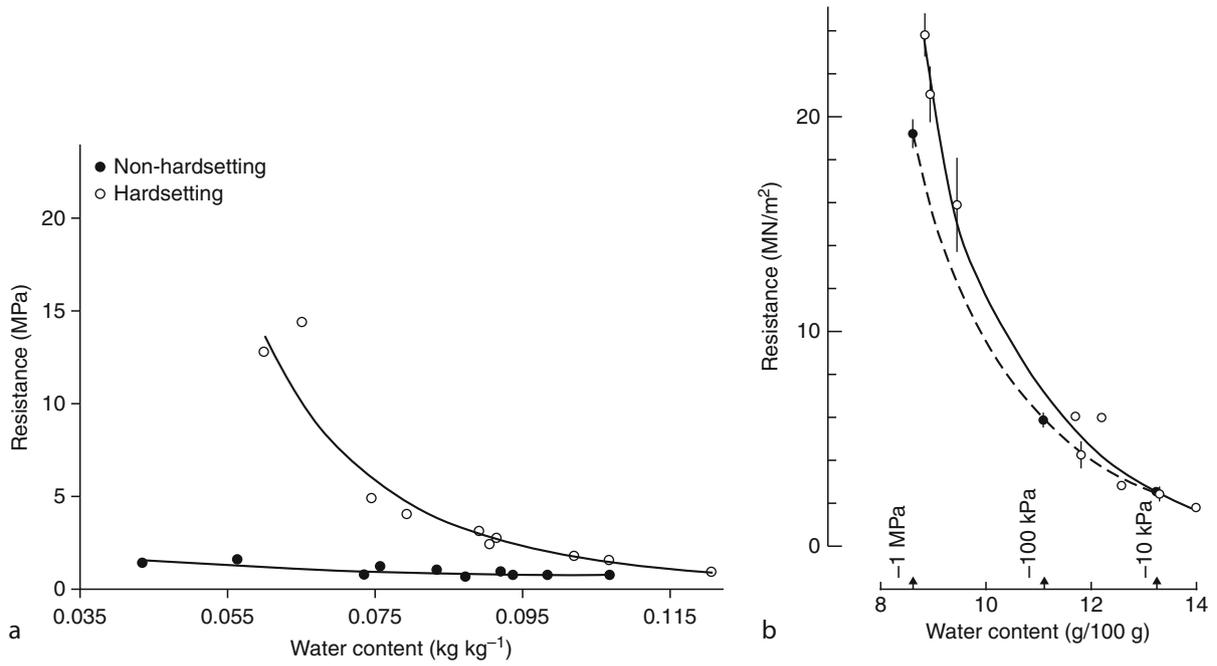
Once wet, the unstable hardsetting soil structure collapses and then shrinks as it dries. This leads to a “massive” soil layer with little or no cracks and greatly reduced pore space (Lima et al., 2006). This hard-set “massive” structure is associated with poor infiltration, a low water holding capacity, and a high soil strength (Figure 1). In many instances, this causes patchy establishment and poor crop and pasture growth. Naturally hardsetting soils are unable to develop water-stable aggregates (Mullins et al., 1990). This means that during wetting, soil aggregates start to swell and become soft. This occurs prior to “slumping” (also referred to as “slaking”) when the aggregates collapse and disintegrate.

Hardsetting can also occur in soil with a high exchangeable sodium percentage (ESP) through the “dispersion” of soil aggregates. This results in clay and silt becoming suspended in the soil solution and causing a breakdown of aggregates (Mullins et al., 1990). Other factors that influence the dispersion of soil aggregates include the soil electrical conductivity (EC), calcium/magnesium ratios, and the organic matter content. Soil types more prone to soil structure decline are sandy loams to clay loams (between 10% to 35% clay), particularly those low in organic matter (<2%) (Giarola et al., 2001).

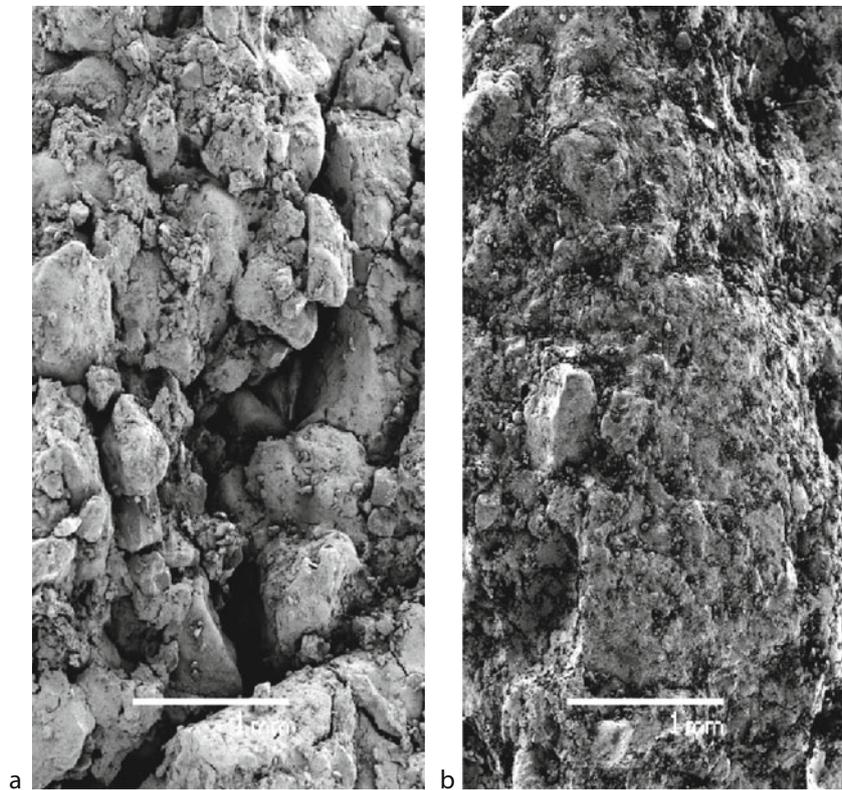
The soil resistance to penetration curve (RP) can be used to differentiate hardsetting from non-hardsetting soils. In soils with hardsetting horizons, in the same moisture range, the variation of soil resistance is much greater than in soils with stable structures (Figure 2). The soil resistance normally exceeds 3 MPa before the soil reaches the permanent wilting point (1,500 kPa of



**Hardsetting Soils: Physical Properties, Figure 1** Soil strength (SS) versus effective stress ( $\sigma'$ ) for non-hardsetting (A1) and hardsetting (AB1) horizons.  $B_d$  is the bulk density.



**Hardsetting Soils: Physical Properties, Figure 2** Curve of soil resistance to penetration for hardsetting and non-hardsetting soils from Brazil (a) and Australia (b). Source: adapted from (a) Giarola et al. (2001) and (b) Mullins et al. (1987).



**Hardsetting Soils: Physical Properties, Figure 3** Structural arrangement of soil particles of (a) non-hardsetting and (b) hardsetting horizons from Cruz das Almas, Bahia, Brazil.

matrix potential). Some soils studied by Mullins et al. (1987) developed a resistance to penetration greater than  $3 \text{ MN m}^{-2}$ , before being dried to a potential of 100 kPa. Ley et al. (1995) found an RP equal to or greater than 2 MPa in some soils from Nigeria, when these soils were dried at a matrix potential of only 100 kPa. Similar results have been obtained for hardsetting soils from the United Kingdom (Young et al., 1991), Australia (Mullins et al. 1987), Tanzania (Mullins, 1997), and Brazil (Fabiola et al., 2003).

In addition to their high cohesion, the denser layers have higher bulk densities ( $1.6\text{--}1.8$  compared to  $1.4\text{--}1.5 \text{ Mg m}^{-3}$ ) and lower permeabilities compared to the softer upper layers (Fabiola et al., 2003; Lima et al., 2006) (Figure 3).

The decrease in total pore volume is another negative consequence of hardsetting behavior, as it affects the biological activity, the movement and capacity of water retention, and the availability of water for plants. The lower pore volume shows a marked effect on the increase of RP during soil drying, which can vary from close to zero to 25 MPa at the point of permanent wilt (matrix potential  $[\psi_m] = 1.5 \text{ MPa}$ ). Values of  $\text{RP} = 3 \text{ MPa}$  were also obtained for a dampness close to  $0.15 \text{ cm}^3 \text{ cm}^{-3}$ , which is sufficient to impede plant growth or emergence (Mullins, 1997).

The tensile strength (TS) of aggregates is another parameter used to recognize hardsetting behavior. Values of  $\text{TS} = 200 \text{ kPa}$  were registered in materials from Australian hardsetting soils after air drying (Ley et al., 1989; Gusli et al., 1994). In Brazil, the TS varied from 37 to 76 kPa in hardsetting horizons with a loamy-sandy-clay texture (Fabiola et al., 2003; Lima et al., 2006).

## Conclusions

Hardsetting soils are structurally unstable soils common in Oceania, Africa, Asia, and South America. Because of their instability to wetting, cultivated hardsetting soils become almost homogenous masses upon drying and present physical problems such as high soil strength, poor infiltration, and crusting, which tend to adversely affect crop performance and management. The latter includes losses in the timeliness of cultivation, as well as a requirement for more frequent irrigation and tillage, leading to further deterioration in soil structure. The lack of defined parameters that indicate the presence of hardsetting behavior and the different degrees of cohesion make it impossible to accurately and easily recognize this behavior in soils.

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## Cross-references

[Aeration of Soils and Plants](#)  
[Compaction of Soil](#)  
[Conditioners, Effect on Soil Physical Properties](#)  
[Crop Emergence, the Impact of Mechanical Impedance](#)  
[Crop Responses to Soil Physical Conditions](#)  
[Infiltration in Soils](#)  
[Layered Soils, Water and Solute Transport](#)  
[Root Responses to Soil Physical Limitations](#)  
[Soil Penetrometers and Penetrability](#)  
[Soil Surface Sealing and Crusting](#)  
[Subsoil Compaction](#)

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## HARVEST TECHNOLOGY

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See *Mechanical Impacts at Harvest and After Harvest Technologies*

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## HEAT ADVECTION

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See *Energy Balance of Ecosystems*

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## HEAT BALANCE

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$R_n + M = C + \lambda E + G$ , where:  $R_n$  - net gain of heat from radiation,  $M$  - net gain of heat from metabolism,  $C$  - loss of sensible heat by convection,  $\lambda E$  - loss of sensible heat by evaporation,  $G$  - conductivity to the environment.

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### Cross-references

[Energy Balance of Ecosystems](#)

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## HEAT CAPACITY

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### Synonyms

Thermal capacity

The quantity of heat required to raise a unit volume of the substance 1 degree of temperature.

### Cross-references

[Coupled Heat and Water Transfer in Soil Thermal Technologies in Food Processing](#)

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## HEAT DIFFUSION

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See *Diffusion in Soils*

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## HEAT OF CONDENSATION

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The amount of heat released when a vapor changes state to a liquid.

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## HEAT OF SUBLIMATION

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The amount of energy required to convert ice directly to a vapor.

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## HEAT OF VAPORIZATION

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The amount of heat required to change a volume of liquid to a vapor.

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## HEAT OF WETTING

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The heat released by a unit mass of initially dry soil when immersed in water. It is related to the soil's specific surface (i.e., the content and composition of the clay fraction).

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## HENRY'S LAW

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The weight of any gas that will dissolve in a given volume of a liquid at constant temperature is directly proportional to the pressure that the gas exerts above the liquid.

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## HOOKE'S LAW

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The deformation (strain) of a body under stress is proportional to the stress applied to it. This law pertains to elastic bodies. The constant of proportionality between stress and strain is known as "Young's modulus."

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Introduction to Environmental Soil Physics (First Edition) 2003 Elsevier Inc. Daniel Hillel (ed.) <http://www.sciencedirect.com/science/book/9780123486554>

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## HORTICULTURE SUBSTRATES, STRUCTURE AND PHYSICAL PROPERTIES

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### Definition

*Horticultural substrate.* It is the life environment of the plant roots, isolated from the parent rock.

*Structure.* Form of spatial arrangement of the solid phase. Structural elements in the horticultural substrate are the primary particles of the solid phase, their complexes, or aggregates and pores, or the space between the solid phase particles and aggregates.

*Physical property.* Physical property is the attribute of a substance that can be observed and measured without changing one substance into another. The main physical parameters characterizing the physical condition of the

horticultural substrate are bulk density, total porosity, container water capacity, available water retention, air capacity, water, and air permeability.

### Introduction

In the second half of the twentieth century, horticultural substrates made mainly from white peat were the most widely used. Now, the assortment of applied materials has increased considerably. Besides the commonly used organic substrates, some other materials are also being used. It is, first of all, the rockwool, produced from the melted diabase or basalt with addition of some dolomite as well as the artificial substrates such as superabsorbents. The physical condition of the applied materials definitely decides on the success of the cultivation. The suitable growth of the plants and their development can be assured, above all, by the proper proportions of the amount of water and air in the substrate (Verdonck et al., 1983). A very important problem in case of the determination of the physical properties of the horticultural substrates is the application of such standard methods, which make possible to receive the comparable results of the measurements (Gabriëls and Verdonck, 1991; Bohne and Günther, 1997).

### Classification of the substrates

Taking into the consideration the materials, which can be used as substrates in horticulture, one can classify them, first of all, into the unary and multicomponent substrates. Among the unary substrates, one discerns the organic, mineral, and artificial ones. To the organic substrates belong white peat, black peat, brown coal, straw, coconut fiber and dust, tobacco, and wood waste. Gravel, sand, grit, keramsite, perlite, and rockwool are grouped among the mineral substrates, while the artificial substrates comprise the phenolic ones, the polyurethane, polyethylene, and polyvinylchloride foams as well as the superabsorbents composed of polyvinyl alcohol, polyoxyethylene, or polyacrylates. To the multicomponent substrates belong the traditional horticultural substrates produced from leaves, sod, heather, compost and garden soil; peat substrates, and standard soils. Standard soils are the substrates, prepared from the materials of defined properties and constant composition. To some of the first such universal horticultural substrates belong John Innes Composts, elaborated in the mid-thirties of the twentieth century in Great Britain. They are formed of the loam, peat, and grit-sand, completed by the addition of some mineral fertilizers. A well-known standard soil is "Einheitserde" elaborated by Anton Frühstorfer in Germany. It is made of white peat, black peat, and loam or clay (Turski et al., 1980).

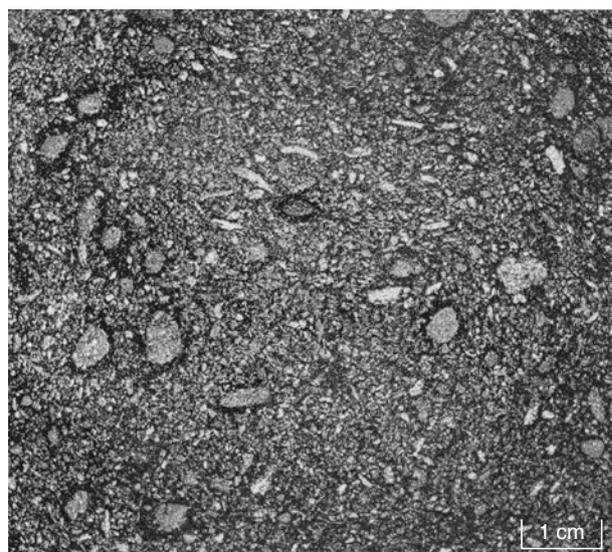
### Structure

Structure considerably decides on such conditions of the plant growth and development, as supply in air and water as well as on the temperature in the root area. In the field conditions, the most suitable physical condition of the soil is guaranteed by the aggregate structure. It is characteristic

for the soils, in which there occur the aggregates – the clumps of the particles of the solid substance only in certain points loosely connected. There is air in the large interaggregate spaces, while the small internal pores of the aggregates keep water in them; this ensures that the plant roots have free access both to the water and to the oxygen. In the horticultural substrates, the aggregate structure is not as much essential as in the natural soils. The amount of water and air in the substrate highly depends on the way of hydration, on the regulating the water outflow, on the dimensions, and on the shape of the container, and not on the solid phase geometry, as it takes place in soil (Fonteno, 1989; Argo, 1998). Very good conditions of the growth and development of the plants, in spite of the lack of aggregate structure, is assured by the rockwool, in which the pressed concentrations of the fibers shape a characteristic sponge-like structure. In the horticultural substrates the aggregate structure occurs most often in case of the loosely heaped-up materials of a considerable contribution of the organic substance, while the structure of separated particles is characteristic for the mineral horticultural substrates (Słowińska-Jurkiewicz and Jaroszuk-Sierocińska, 2007). Very advantageous structure of horticultural substrates represents Figure 1.

### Physical properties

Traditionally, such substrates were considered to be the most suitable for the cultivation of the garden plants in which one-half of the volume is occupied by the solid phase, and the other by the pores, just as in the mineral



**Horticulture Substrates, Structure and Physical Properties, Figure 1** Structure of the mixture of white peat (50%, v/v) with coconut fiber (50%, v/v). Image in 256 gray degrees of polished opaque block (surface dimensions 8 × 9 cm) developed from this substrate impregnated with polyester resin. Color of the pores is *black* and of the solid phase – *gray* (Słowińska-Jurkiewicz and Jaroszuk-Sierocińska, 2007).

soils (Penningsfeld and Kurzman, 1966). De Boodt (1965) stated that an ideal substrate should be characterized by a considerably larger total porosity, about  $0.85 \text{ m}^3 \text{ m}^{-3}$  and a low bulk density,  $0.215 \text{ Mg m}^{-3}$ . Such conditions can be realized, first of all, in the soilless substrates, produced on the base of peat, as well as in the modern substrates, such as rockwool. Pores and their dimensions play an important part in the water and air conditions. In the pores there is either the water or the air. One admits that the large pores contain the air (except the situation of a complete saturation of substrate with water), while the small pores are filled with water. Drzal et al. (1999) introduced a classification on large pores (macropores), of dimensions  $>416 \mu\text{m}$ , from which the water flows out, quickly, under the influence of the gravitation force, middle-size pores (mesopores), of dimension from  $416$  up to  $10 \mu\text{m}$ , in which the water, available for the plants, is retained against the gravitation, as well as the small pores (micropores), covering the range from  $10$  up to  $0.2 \mu\text{m}$ . The micropores contain the water which is not used by the plants in case of a normal hydration, and being a reserve in the situation of a water stress. The pores of dimensions below  $0.2 \mu\text{m}$  or the ultramicropores, keep the water unavailable for plants. To the dimension of pores  $416 \mu\text{m}$  corresponds the water potential of  $-0.7 \text{ kPa}$ , to the dimension  $10 \mu\text{m}$ , water potential  $-31 \text{ kPa}$ , and to dimension of  $0.2 \mu\text{m}$ , water potential  $-1.5 \text{ MPa}$ . According to White and Mastalerz (1966), De Boodt and Verdonck (1972), and Fonteno (1989, 1993) in case of characterizing the water properties of the horticultural substrates, as a basic parameter there should be named the container water capacity, defined as the amount of water, remained in the substrate after the free outflow of the gravitational water, but before the beginning of evaporation. This amount depends not only on the character of the substrate but considerably on the dimensions, and also on the shape of the container, in which the plant is

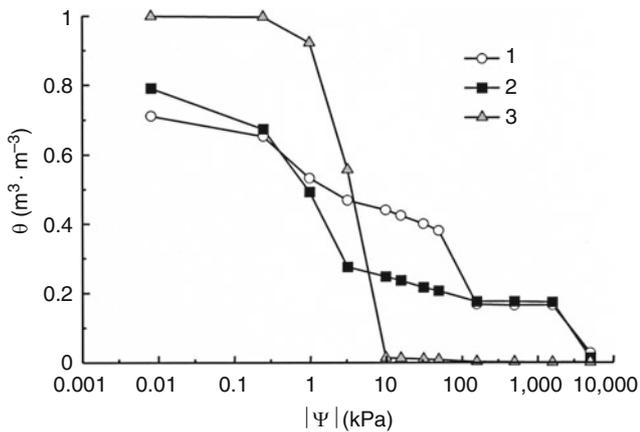
cultivated (Drzal et al., 1999). After the irrigation and outflow, the level of free water occurs on the bottom of the container. For every 1-cm increase of the height above the bottom of the container the water potential decreases by  $0.1 \text{ kPa}$ , and, this way, decreases the possibility of its keeping. For the container of 20-cm high, the average water potential corresponding to the container water capacity is equal to  $-1 \text{ kPa}$ . In regard to the water retention in the substrates, De Boodt and Verdonck (1972) applied the concept of an easily available water in the range of water potential from  $-1$  to  $-5 \text{ kPa}$  and water buffering capacity from  $-5$  to  $-10 \text{ kPa}$ . Brückner (1997) made a difference between the light available water retention in the range of water potential from  $-1$  to  $-10 \text{ kPa}$  and the heavy available water retention in the range from  $-10 \text{ kPa}$  to  $-1.5 \text{ MPa}$ . Light available water retention is especially important for steering the irrigation. It should begin soon after its consumption by the plants, and thus in the case of water potential being  $-10 \text{ kPa}$ . Aside of the characteristics determining a capability of the material to collecting the water, very important are the parameters determining its capability to water filtration, both in the saturated and in the unsaturated zone (Sławiński et al., 1996). In the condition of the saturation the movement of water is determined by the large pores. With the moisture decrease in the substrate, after the water outflow from the large pores, the movement of water takes place in the smaller pores also, which results in a more tortuous route of water outflow (Fonteno, 1993). In the substrate environment, aside of water, also the air plays an important part (Caron and Nkongolo, 1999). In an ideal environment of root growth of porosity  $0.85 \text{ m}^3 \text{ m}^{-3}$  placed in the pot of 15-cm high, in state of container water capacity, the air should occupy  $0.25 \text{ m}^3 \text{ m}^{-3}$  and water,  $0.60 \text{ m}^3 \text{ m}^{-3}$  (De Boodt and Verdonck, 1972). It should be remembered that the change of the bulk density of substrates, related to their compaction during the transport

**Horticulture Substrates, Structure and Physical Properties, Table 1** Basic physical properties of loose horticultural substrates

Type of substrate	Bulk density ( $\text{Mg m}^{-3}$ )	Total porosity <sup>a</sup> ( $\text{m}^3 \text{ m}^{-3}$ )	Container water capacity at $-1 \text{ kPa}$		Available water retention from $-1$ to $-10 \text{ kPa}$		Air capacity at $-1 \text{ kPa}$ <sup>b</sup> ( $\text{m}^3 \text{ m}^{-3}$ )
			( $\text{kg kg}^{-1}$ )	( $\text{m}^3 \text{ m}^{-3}$ )	( $\text{kg kg}^{-1}$ )	( $\text{m}^3 \text{ m}^{-3}$ )	
Wheat peat	0.127	0.911	4.193	0.536	0.718	0.090	0.375
Peat substrate	0.247	0.853	3.211	0.796	0.962	0.238	0.057
Soil with coconut fiber	0.207	0.862	2.960	0.614	0.629	0.129	0.248
Coconut fiber	0.053	0.971	9.290	0.501	4.627	0.250	0.470
Composting bark	0.198	0.874	2.806	0.552	1.190	0.236	0.322
Pine bark	0.143	0.900	1.429	0.204	0.327	0.047	0.696
Sand	1.441	0.453	0.200	0.284	0.096	0.138	0.169
Grit	1.485	0.481	0.046	0.068	0.032	0.047	0.413
Keramsite	0.702	0.710	0.337	0.237	0.024	0.017	0.473
Perlite	0.156	0.943	2.620	0.409	1.166	0.182	0.534
Rockwool	0.082	0.971	11.239	0.922	11.053	0.906	0.049

<sup>a</sup>Total porosity calculated according to the values of particle density and bulk density

<sup>b</sup>Air capacity at  $-1 \text{ kPa}$  calculated as a difference between the total porosity and the container water capacity value



**Horticulture Substrates, Structure and Physical Properties, Figure 2** Water retention curves of three horticultural substrates: white peat – 1, coconut fiber – 2, and rockwool – 3. On the horizontal axis there are the absolute values of water potential ( $|\Psi|$ ) in kPa and on vertical axis there are the values of moisture content ( $\theta$ ) in  $\text{m}^3 \text{m}^{-3}$  (Jaroszuk-Sierocińska and Słowińska-Jurkiewicz, 2009).

and performing various cultivating works, can result in a radical decrease in the air capacity (Brückner, 1997; Jaroszuk and Słowińska-Jurkiewicz, 2003). The values of the basic physical properties of most often used substrates are listed in Table 1 (Jaroszuk and Słowińska-Jurkiewicz, 2005).

## Conclusions

Horticultural substrates show the most various physical properties, depending on the character of the materials used to their production. Among the actually used substrates, the best physical condition, from the point of view of the horticultural production, is characteristic for rockwool and coconut fiber (Figure 2). These substrates can certainly substitute the white peat in the process of horticultural production, what surely results in protections of the bogs under the menace of the excessive exploitation.

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## Cross-references

- [Pore Size Distribution](#)  
[Soil Water Management](#)

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## HUMUS

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The well decomposed, more or less stable part of the organic matter in mineral soils.

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## HYDRAULIC DIFFUSIVITY

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The ratio between the flux of water and the gradient of soil wetness. This term is somewhat misleading, since it does not refer to diffusion as such but to convection. The term is taken from the analogy to the diffusion equation (Fick's law), stating that the rate of diffusion is proportional to the concentration gradient.

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## HYDRAULIC GRADIENT

The slope of the hydraulic grade line which indicates the change in pressure head per unit of distance.

## HYDRAULIC HEAD

The sum of the pressure head (hydrostatic pressure relative to atmospheric pressure) and the gravitational head (elevation relative to a reference level). The gradient of the hydraulic head is the driving force for water flow in porous media.

## Bibliography

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## HYDRAULIC PROPERTIES OF UNSATURATED SOILS

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### Definition

*Hydraulic Properties of Unsaturated Soils.* Properties reflecting the ability of a soil to retain or transmit water and its dissolved constituents.

### Introduction

Many agrophysical applications require knowledge of the hydraulic properties of unsaturated soils. These properties reflect the ability of a soil to retain or transmit water and its dissolved constituents. For example, they affect the partitioning of rainfall and irrigation water into infiltration and runoff at the soil surface, the rate and amount of redistribution of water in a soil profile, available water in the soil root zone, and recharge to or capillary rise from the groundwater table, among many other processes in the unsaturated or vadose zone between the soil surface and the groundwater table. The hydraulic properties are also critical components of mathematical models for

studying or predicting site-specific water flow and solute transport processes in the subsurface. This includes using models as tools for designing, testing, or implementing soil, water, and crop management practices that optimize water use efficiency and minimize soil and water pollution by agricultural and other contaminants. Models are equally needed for designing or remediating industrial waste disposal sites and landfills, or assessing the for long-term stewardship of nuclear waste repositories.

Predictive models for flow in variably saturated soils are generally based on the Richards equation, which combines the Darcy–Buckingham equation for the fluid flux with a mass conservation equation to give (Richards, 1931):

$$\frac{\partial \theta(h)}{\partial t} = \frac{\partial}{\partial z} \left[ K(h) \frac{\partial h}{\partial z} - K(h) \right] \quad (1)$$

in which  $\theta$  is the volumetric water content ( $L^3 L^{-3}$ ),  $h$  is the pressure head (L),  $t$  is time (T),  $z$  is soil depth (positive down), and  $K$  is the hydraulic conductivity ( $L T^{-1}$ ). Equation 1 holds for one-dimensional vertical flow; similar equations can be formulated for multidimensional flow problems. The Richards equation contains two constitutive relationships, the soil water retention curve,  $\theta(h)$ , and the unsaturated soil hydraulic conductivity function,  $K(h)$ . These hydraulic functions are both strongly nonlinear functions of  $h$ . They are discussed in detail below.

### Water retention function

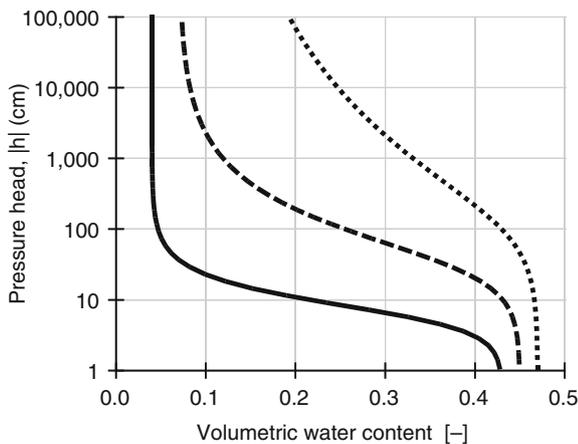
The soil water retention curve,  $\theta(h)$ , describes the relationship between the water content,  $\theta$ , and the energy status of water at a given location in the soil. Many other names may be found in the literature, including soil moisture characteristic curve, the capillary pressure–saturation relationship, and the pF curve. The retention curve historically was often given in terms of pF, which is defined as the negative logarithm (base 10) of the absolute value of the pressure head measured in centimeters. In the unsaturated zone, water is subject to both capillary forces in soil pores and adsorption onto solid phase surfaces. This leads to negative values of the pressure head (or matric head) relative to free water, or a positive suction or tension. As opposed to unsaturated soils, the pressure head  $h$  is positive in a saturated system. More formally, the pressure head is defined as the difference between the pressures of the air phase and the liquid phase. Capillary forces are the result of a complex set of interactions between the solid and liquid phases involving the surface tension of the liquid phase, the contact angle between the solid and liquid phases, and the diameter of pores.

Knowledge of  $\theta(h)$  is essential for the hydraulic characterization of a soil, since it relates an energy density (potential) to a capacity (water content). Rather than using the pressure head (energy per unit weight of water), many agrophysical applications use the pressure or matric potential (energy per unit volume of water, usually measured in Pascal, Pa),  $\psi_m = \rho_w g h$ , where  $\rho_w$  is the density of water ( $ML^{-3}$ ) and  $g$  the acceleration of gravity ( $LT^{-2}$ ).

Figure 1 shows typical soil water retention curves for relatively coarse-textured (e.g., sand and loamy sand), medium-textured (e.g., loam and sandy loam), and fine-textured (e.g., clay loam, silty loam, and clay) soils. The curves in Figure 1 may be interpreted as showing the equilibrium water content distribution above a relatively deep water table where the pressure head is zero and the soil fully saturated. The plots in Figure 1 show that coarse-textured soils lose their water relatively quickly (at small negative pressure heads) and abruptly above the water table, while fine-textured soils lose their water much more gradually. This reflects the particle or pore-size distribution of the medium involved. While the majority of pores in coarse-textured soils have larger diameters and thus drain at relatively small negative pressures, the majority of pores in fine-textured soils do not drain until very large tensions (negative pressures) are applied.

As indicated by the plots in Figure 1, the water content varies between some maximum value, the saturated water content,  $\theta_s$ , and some small value, often referred to as the residual (or irreducible) water content,  $\theta_r$ . As a first approximation and on intuitive ground, the saturated water content is equal to the porosity, and  $\theta_r$  equal to zero. In reality, however, the saturated water content,  $\theta_s$ , of soils is generally smaller than the porosity because of entrapped and dissolved air. The residual water content  $\theta_r$  is likely to be larger than zero, especially for fine-textured soils with their large surface areas, because of the presence of adsorbed water. Most often  $\theta_s$  and especially  $\theta_r$  are treated as fitting parameters without much physical significance.

Soil water retention curves such as shown in Figure 1 are not unique but depend on the history of wetting and drying. Most often, the soil water retention curve is determined by gradually desaturating an initially saturated soil by applying increasingly higher suctions, thus producing a main drying curve. One could similarly slowly wet an



**Hydraulic Properties of Unsaturated Soils, Figure 1** Typical soil water retention curves for relatively coarse- (solid line), medium- (dashed line), and fine-textured (dotted line) soils. The curves were obtained using Equation 6a assuming hydraulic parameter values as listed in Table 1.

initially very dry sample to produce the main wetting curve, which is generally displaced by a factor of 1.5–2.0 toward higher pressure heads closer to saturation. This phenomenon of having different wetting and drying curves, including primary and secondary scanning curves is referred to hysteresis. Hysteresis is caused by the fact that drainage is determined mostly by the smaller pore in a certain pore sequence, and wetting by the larger pores (this effect is often referred to as the ink bottle effect). Other factors contributing to hysteresis are the presence of different liquid–solid contact angles for advancing and receding water menisci, air entrapment during wetting, and possible shrink–swell phenomena of some soils.

### Hydraulic conductivity function

The hydraulic conductivity characterizes the ability of a soil to transmit water. Its value depends on many factors such as the pore-size distribution of the medium, and the tortuosity, shape, roughness, and degree of interconnectedness of the pores. The hydraulic conductivity decreases considerably as soil becomes unsaturated since less pore space is filled with water, the flow paths become increasingly tortuous, and drag forces between the fluid and the solid phases increase.

The unsaturated hydraulic conductivity function gives the dependency of the hydraulic conductivity on the water content,  $K(\theta)$ , or pressure head,  $K(h)$ . Figure 2 presents examples of typical  $K(\theta)$  and  $K(h)$  functions for relatively coarse-, medium-, and fine-textured soils. Notice that the hydraulic conductivity at saturation is significantly larger for coarse-textured soils than fine-textured soils. This difference is often several orders of magnitude. Also notice that the hydraulic conductivity decreases very significantly as the soil becomes unsaturated. This decrease, when expressed as a function of the pressure head (Figure 2; right), is much more dramatic for the coarse-textured soils. The decrease for coarse-textured soils is so large that at a certain pressure head the hydraulic conductivity becomes smaller than the conductivity of the fine-textured soil. The water content where the conductivity asymptotically becomes zero (Figure 2; left) is often used as an alternative working definition for the residual water content,  $\theta_r$ .

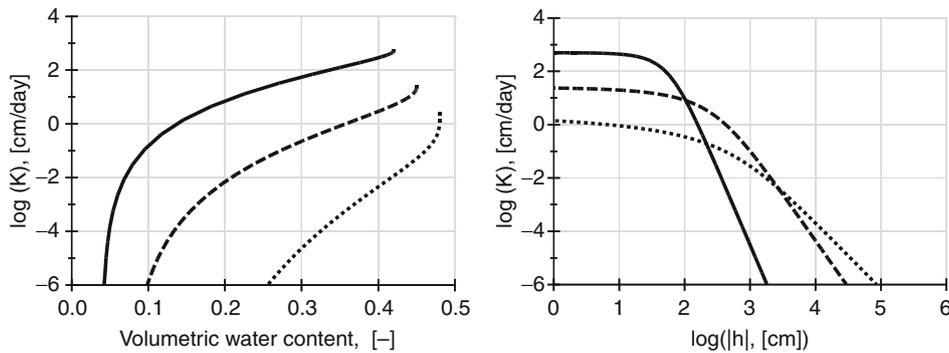
### Soil water diffusivity

Another hydraulic function often used in theoretical and management application of unsaturated flow theories is the soil water diffusivity,  $D(\theta)$ , ( $L^2 T^{-1}$ ), which is defined as

$$D(\theta) = K(\theta) \left| \frac{dh}{d\theta} \right|. \quad (2)$$

This function appears when Equation 1 is transformed into a water-content-based equation in which  $\theta$  is now the dependent variable:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left[ D(\theta) \frac{\partial h}{\partial z} - K(\theta) \right]. \quad (3)$$



**Hydraulic Properties of Unsaturated Soils, Figure 2** Typical curves of the hydraulic conductivity  $K$ , as a function of the pressure head (left) and water content (right) for coarse- (solid line), medium- (dashed line), and fine-textured (dotted line) soils. The curves were obtained using Equation 6b assuming hydraulic parameter values as listed in Table 1.

Equation 3 is very attractive for approximate analytical modeling of unsaturated flow processes, especially for modeling horizontal (without the  $K(\theta)$  gravity term) and vertical infiltration (e.g., Philip, 1969; Parlange, 1980). However, the water-content-based equation is less attractive for more comprehensive numerical modeling of flow in layered media, flow in media that are partially saturated and partially unsaturated, and for highly transient flow problems.

### Analytical representations

To enable their use in analytical or numerical models for unsaturated flow, the soil hydraulic properties are often expressed in terms of simplified analytical expressions. A large number of functions have been proposed over the years to describe the soil water retention curve,  $\theta(h)$ , and the hydraulic conductivity function,  $K(h)$  or  $K(\theta)$ . A comprehensive review of the performance of some of many these models is given by Leij et al. (1997). The functions range from completely empirical equations to models based on the simplified conceptual picture that soils are made up of a bundle of equivalent capillary tubes that contain and transmit water.

While extremely simplistic as indicated by Tuller and Or (2001) among others, conceptual models that view a soil as a bundle of capillaries of different radii are still useful for explaining the shape of the water retention curve for different textures, as well as to provide a means for predicting the hydraulic conductivity function from soil water retention information. These models typically assume that pores at a given pressure head are either completely filled with water, or empty, depending upon the applied suction. Flow in each water-filled capillary tube is subsequently calculated using Poiseuille's law for flow in cylindrical pores. By adding the contribution of all capillaries that are still filled with water at a particular pressure head, making some assumption about how small and large capillaries connect to each other in sequence

(using a cut-and-paste concept of a cross-section of the medium containing different-sized pores), and then integrating over all water-filled capillaries leads to the hydraulic conductivity of the complete set of capillaries, and consequently of the soil itself. The approach allows information of the soil water retention curve to be translated in predictive equations for the unsaturated hydraulic conductivity. Many theories of this type, often referred to also as statistical pore-size distribution models, have been proposed in the past, including Childs and Collis-George (1950), Burdine (1953), Millington and Quirk (1961), and Mualem (1976). A review of the different approaches is given by Mualem (1992). Examples of analytical  $\theta(h)$  and  $K(h)$  equations resulting from this approach are the hydraulic functions of Brooks and Corey (1964), based on the approach by Burdine (1953), and equations by van Genuchten (1980) and Kosugi (1996), based on the theory of Mualem (1976).

The classical equations of Brooks and Corey (1964) for  $\theta(h)$ ,  $K(h)$ , and  $D(\theta)$  are given by

$$\theta = \begin{cases} \theta_r + (\theta_s - \theta_r) \left| \frac{h_e}{h} \right|^\lambda & h < h_e \\ \theta_s & h \geq h_e \end{cases} \quad (4a)$$

$$K(h) = K_s S_e^{2/\lambda + 1} \quad (4b)$$

$$D(\theta) = \frac{K_s}{\alpha(\theta_s - \theta_r)} S_e^{1/\lambda + 1} \quad (4c)$$

where, as before,  $\theta_r$  is the residual water content ( $L^3 L^{-3}$ ),  $\theta_s$  is the saturated water content ( $L^3 L^{-3}$ ),  $h_e$  is often referred to as the air-entry value (L),  $\lambda$  is a pore-size distribution index characterizing the width of the soil pore-size distribution,  $K_s$  is the saturated hydraulic conductivity ( $LT^{-1}$ ),  $l$  a pore-connectivity parameter assumed to be 2.0 in the original study of Brooks and Corey (1964), and  $S_e = S_e(h)$  is effective saturation given by

$$S_e(h) = \frac{\theta(h) - \theta_r}{\theta_s - \theta_r} \quad (5)$$

For completeness we have given here also the expression for the soil water diffusivity,  $D(\theta)$ . Note that Equations 4b and 4c contain parameters that are also present in Equation 4a, in particular  $\theta_r$  and  $\theta_s$  through Equation 5, as well as  $h_e$  and  $\lambda$ . The value of  $\lambda$  in Equation 4a reflects the steepness of the retention function and is relatively large for soils with a relatively uniform pore-size distribution (generally coarse-textured soils such as those shown in Figures 1 and 2), but small for soils having a wide range of pore sizes.

One property of Equation 4a is the presence of a sharp break in the retention curve at the air-entry value,  $h_e$ . This break (or discontinuity in the slope of the function) is often visible in retention data for coarse-textured soils, but may not be realistic for fine-textured soils and soils having a relatively broad pore- or particle-size distribution. A sharp break is similarly present in the hydraulic conductivity function when plotted as a function of the pressure head, but not versus the water content. As an alternative, van Genuchten (1980) proposed a set of equations that exhibit a more smooth sigmoidal shape. The van Genuchten equations for  $\theta(h)$ ,  $K(h)$ , and  $D(\theta)$  are given by:

$$\theta(h) = \theta_r + \frac{\theta_s - \theta_r}{(1 + |\alpha h|^n)^m} \quad (m = 1 - 1/n; n > 1) \quad (6a)$$

$$K(h) = K_s S_e^l \left[ 1 - \left( 1 - S_e^{1/m} \right)^m \right]^2 \quad (6b)$$

$$D(\theta) = \frac{(1-m)K_s}{\alpha m(\theta_s - \theta_r)} S_e^{l-1/m} \left[ \left( 1 - S_e^{1/m} \right)^{-m} + \left( 1 - S_e^{1/m} \right)^m - 2 \right] \quad (6c)$$

respectively, where  $\alpha$  ( $L^{-1}$ ),  $n$  ( $-$ ), and  $m$  ( $= 1 - 1/n$ ) ( $-$ ) are shape parameters, and  $l$  is the pore-connectivity parameter ( $-$ ). The parameter  $n$  in Equation 6 tends to be large for soils with a relatively uniform pore-size distribution and small for soils having a wide range of pore sizes. The pore-connectivity parameter  $l$  in Equation 6b was estimated by Mualem (1976) to be about 0.5 as an average for many soils. However, many other values

for  $l$  have been suggested in various studies. Based on an analysis of a large data set from the UNSODA database, Schaap and Leij (2000) recommended using  $l$  equal to  $-1$  as a more appropriate value for most soil textures.

Equations 6a, 6b, and 6c assume the restrictive relationship  $m = 1 - 1/n$ , which simplifies the predictive  $K(h)$  expression compared to leaving  $m$  and  $n$  as independent parameters in Equation 6b. In particular, the convex and concave curvatures at the high and low pressure heads in Figure 1 have then a particular relationship with each other. Other restrictions on Equation 6a have been used also. For example, Haverkamp et al. (2005) used the restriction  $m = 1 - 2/n$  in connection with Equation 6a and Burdine's (1953) model to produce a different expression for  $K(h)$ . The restrictions are not formally needed, since they limit the flexibility of Equation 6a in describing experimental data. However, the predicted  $K(h)$  function obtained with the theories of Burdine or Mualem becomes then extremely complicated by containing incomplete beta or hypergeometric functions, thus limiting the practicality of the analytical functions.

Rawls et al. (1982) provided average values of the parameters in the Brooks and Corey (1964) soil hydraulic parameters for 11 soil textural classes of the U.S. Department of Agriculture (USDA) textural triangle. Carsel and Parrish (1988) gave similar values for the van Genuchten (1980) parameters for 12 USDA soil textural classes. In Table 1, we list typical van Genuchten hydraulic parameter values for relative coarse-, medium-, and fine-textured soils. The data in this table were actually used to calculate the water retention and hydraulic conductivity functions, shown in Figures 1 and 2, respectively, with Equations 6a, b. Average values such as those given in Table 1, or provided in more detail by Rawls et al. (1982) and Carsel and Parrish (1988), are often referred to as textural class averaged pedotransfer functions. Pedotransfer functions are relationships that use more easily measured of readily available soil data to estimate the unsaturated soil hydraulic parameters or properties (Bouma and van Lanen, 1987; Leij et al., 2002; Pachepsky and Rawls, 2004).

We note that Equations 4 and 6 provide only two examples in which the hydraulic properties are described analytically. Many other combinations (Leij et al., 1997; Kosugi et al., 2002) are possible and have been used. For example, the combination of Equation 6a for  $\theta(h)$  with a simple expression like

**Hydraulic Properties of Unsaturated Soils, Table 1** Typical values of the soil hydraulic parameters in the analytical functions of van Genuchten (1980) for relatively coarse-, medium-, and fine-textured soils. The parameters were used to calculate the hydraulic properties plotted in Figures 1 and 2 using Equations 6a and 6b, respectively

Soil texture	$\theta_r$ ( $\text{cm day}^{-1}$ )	$\theta_s$ ( $\text{cm}^3 \text{ cm}^{-3}$ )	$\alpha$ ( $\text{cm}^3 \text{ cm}^{-3}$ )	$n$ ( $\text{cm}^{-1}$ )	$K_s$ ( $-$ )
Coarse	0.045	0.430	0.145	2.68	712.8
Medium	0.057	0.410	0.124	2.28	350.2
Fine	0.020	0.540	0.0010	1.2	45.0

$$K(h) = K_s S_e^b \quad (7)$$

which is essentially identical to Equation 4b, for  $K(h)$  is also very realistic. Another attractive alternative equation for  $K(h)$  is of the form (e.g., Vereecken et al., 1989)

$$K(h) = \frac{K_s}{1 + |ah|^b} \quad (8)$$

Many alternative expressions have been used also for the soil water diffusivity function,  $D(\theta)$ , mostly to facilitate simplified analytical analyses of unsaturated flow problems (e.g., Parlange, 1980).

### Experimental procedures

A large number of experimental techniques can be used to estimate the hydraulic properties of unsaturated soils. A direct approach for the water retention function would be to measure a number of water content ( $\theta$ ) and pressure head ( $h$ ) pairs, and then to fit a particular retention function to the data. Direct measurement techniques include methods using a hanging water column, pressure cells, pressure plate extractors, suction tables, soil freezing, and many other approaches. Comprehensive reviews of various methods are given by Gee and Ward (1999) and Dane and Hopmans (2002). Once the pairs of  $\theta$  and  $h$  data are obtained, the data may be analyzed in terms of specific analytical water retention and conductivity functions such as those discussed earlier. Several convenient software packages are available for this purpose (van Genuchten et al., 1991; Wraith and Or, 1998). Alternatively, the data can be analyzed without assuming specific analytical functions for  $\theta(h)$  and  $K(h)$  or  $K(\theta)$ . This could be done using linear, cubic spline, or other interpolation techniques (Kastanek and Nielsen, 2001; Bitterlich et al., 2004).

Similar direct measurement approaches involving pairs of conductivity (or diffusivity) and pressure head (or water content) data are also possible for the  $K(h)$  and  $D(\theta)$  functions, at least in principle (Dane and Topp, 2002), including for the saturated hydraulic conductivity,  $K_s$ . The saturated hydraulic conductivity can be measured in the laboratory using a variety of constant or falling head methods, and in the field using single or double ring infiltrometers, constant head permeameters, and various auger-hole and piezometer methods (Dane and Topp, 2002). Unfortunately, because of the strongly nonlinear nature of the soil hydraulic properties, pairs for the  $K(h)$  and  $D(\theta)$  data are not easily measured directly, especially at relatively low (negative) pressure heads, unless more specialized techniques are used such as centrifuge methods (Nimmo et al., 2002). Even then, the data are generally not distributed evenly over the entire water content range of interest. Consequently, unsaturated hydraulic conductivity properties are most often estimated using inverse or parameter estimation procedures.

Parameter estimation methods generally involve the measurement during some experiment of one or several

capacity or flow attributes (e.g., water contents, pressure heads, boundary fluxes), which are then used in combination with a mathematical solution (generally numerical) to obtain estimates of the hydraulic parameters such as those that appear in Equations 4 and 6, or other functions. Popular methods include one-step and multi-step outflow methods (Kool et al., 1987; van Dam et al., 1994), tension infiltrometers methods (Šimůnek et al., 1998a), and evaporation methods (Šimůnek et al., 1998b), although many other laboratory and field methods also exist or can be similarly employed (Hopmans et al., 2002). This also pertains to different approaches for minimizing the objective function, including quantification of parameter uncertainty (Abbaspour et al., 2001; Vrugt and Robinson, 2007). Very attractive now also is the use of combined hard (e.g., directly measured) and soft (e.g., indirectly estimated) data, including hydrogeophysical measurements and information derived from pedotransfer functions, to extract the most out of available information (e.g., Kowalski et al., 2004; Segal et al., 2008).

### Hydraulic properties of structured soils

The Richards equation 1 typically predicts a uniform flow process in the vadose zone. Unfortunately, the vadose zone can be extremely heterogeneous at a range of scales, from the microscopic (e.g., pore scale) to the macroscopic (e.g., field or larger scale). Some of these heterogeneities can lead to a preferential (or bypass) flow process that macroscopically is very difficult to capture with the standard Richards equation. One obvious example of preferential flow is the rapid movement of water and dissolved solutes through soil macropores (e.g., between soil aggregates, or created by earthworms or decayed root channels or rock fractures), with much of the water bypassing (short-circuiting) the soil or rock matrix. However, many other causes of preferential flow exist, such as flow instabilities caused by soil textural changes or water repellency (Hendrickx and Flury, 2001; Šimůnek et al., 2003; Ritsema and Dekker, 2005), and lateral funneling of water along sloping soil layers (e.g., Kung, 1990).

While uniform flow in granular soils is traditionally described with a single-porosity model such as the Richards equation given by Equation 1, flow in structured media can be described using a variety of dual-porosity, dual-permeability, multi-porosity, and/or multi-permeability models (Šimůnek and van Genuchten, 2008; Köhne et al., 2009). While single-porosity models assume that a single pore system exists that is fully accessible to both water and solute, dual-porosity and dual-permeability models both assume that the porous medium consists of two interacting pore regions, one associated with the inter-aggregate, macropore, or fracture system, and one comprising the micropores (or intra-aggregate pores) inside soil aggregates or the rock matrix. Whereas dual-porosity models assume that water in the matrix is stagnant, dual-permeability models allow also for water flow within the soil or rock matrix.

To avoid over-parameterization of the governing equations, one useful simplifying approach is to assume instantaneous hydraulic equilibration between the fracture and matrix regions. In that case, the Richards equation can still be used, but now with composite hydraulic properties of the form (e.g., Peters and Klavetter, 1988)

$$\theta(h) = w_f \theta_f(h) + w_m \theta_m(h) \quad (9a)$$

$$K(h) = w_f K_f(h) + w_m K_m(h) \quad (9b)$$

where the subscripts  $f$  and  $m$  refer to the fracture (macropore) and matrix (micropore) regions, respectively, and where  $w_i$  are volumetric weighting factors for the two overlapping regions such that  $w_f + w_m = 1$ . Rather than using Equations 6a,b directly in Equations 9a and 9b, Durner (1994) proposed a slightly different set of equations for the composite functions as follows

$$S_e(h) = \frac{\theta(h) - \theta_r}{\theta_s - \theta_r} = \frac{w_f}{[1 + |\alpha_f h|^{n_f}]^{m_f}} + \frac{w_m}{[1 + |\alpha_m h|^{n_m}]^{m_m}} \quad (10a)$$

$$K(S_e) = \frac{(w_f S_{e_f} + w_m S_{e_m})^l \{w_f \alpha_f [1 - (1 - S_{e_f}^{1/m_f})^{m_f}] + w_m \alpha_m [1 - (1 - S_{e_m}^{1/m_m})^{m_m}]\}^2}{K_s (w_f \alpha_f + w_m \alpha_m)^2} \quad (10b)$$

where  $\alpha_i$ ,  $n_i$ , and  $m_i (=1 - 1/n_i)$  are empirical parameters of the separate hydraulic functions ( $i=f,m$ ). An example of composite retention and hydraulic conductivity functions based on Equations 10a and 10b is shown in Figure 2 for the following set of parameters:  $\theta_r = 0.00$ ,  $\theta_s = 0.50$ ,  $l = 0.5$ ,  $K_s = 1 \text{ cm d}^{-1}$ ,  $\alpha_m = 0.01 \text{ cm}^{-1}$ ,  $n_m = 1.50$ ,  $w_m = 0.975$ ,  $w_f = 0.025$ ,  $\alpha_f = 1.00 \text{ cm}^{-1}$ , and  $n_f = 5.00$ . The fracture domain in this case represents only 2.5% of the entire pore space, but accounts for almost 90% of the hydraulic conductivity close to saturation (Figure 3).

While still leading to uniform flow, models using such composite media properties do allow for faster flow and transport during conditions near saturation, and as such

provide more realistic simulations of field data than the standard approach using unimodal hydraulic properties of the type shown in Figures 1 and 2. In soils, the two parts of the conductivity curves may be associated with soil structure (near saturation) and soil texture (at lower negative pressure heads).

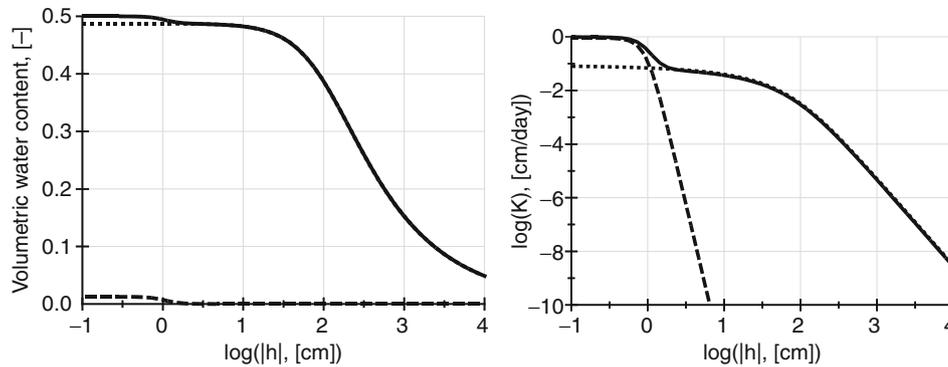
The use of composite hydraulic functions such as those shown in Figure 2 is consistent with field measurements suggesting that the macropore conductivity of soils at saturation is generally about one to two orders of magnitude larger than the matrix conductivity at saturation, depending upon texture. These findings were confirmed by Schaap and van Genuchten (2006) using a detailed neural network analysis of the UNSODA unsaturated soil hydraulic database (Leij et al., 1996). The analysis revealed a relatively sharp decrease in the conductivity away from saturation and a slower decrease afterward. Schaap and van Genuchten (2006) suggested an improved composite function for  $K(h)$  to account for the effects of macropores near saturation as follows:

$$K(h) = \left( \frac{K_s}{K_m(h)} \right)^{R(h)} K_m(h) \quad (11a)$$

where

$$R(h) = \begin{cases} 0 & h < -40 \text{ cm.} \\ 0.2778 + 0.00694h & -40 \leq h < -4 \text{ cm} \\ 1 + 0.1875h & -4 \leq h \leq 0 \text{ cm} \end{cases} \quad (11b)$$

and where  $K_m(h)$  is the traditional hydraulic conductivity function for the matrix as given by Equation 6b. Equations 11a and 11b were found to produce very small systematic errors between the observed (UNSODA) and calculated hydraulic conductivities across a wide range of pressure heads between saturation and  $-150 \text{ m}$ . While the macropore contribution was most significant between pressure heads 0 and  $-4 \text{ cm}$ , its influence on the conductivity function extended to pressure heads as low as  $-40 \text{ cm}$  (Equation 11b).



**Hydraulic Properties of Unsaturated Soils, Figure 3** Bimodal water retention (left) and hydraulic conductivity (right) functions as described with the composite soil hydraulic model of Durner (1994).

### Multiphase constitutive relationships

The use of Equation 1 implies that the air phase has no effect of water flow. This is a realistic assumption for most flow simulations, except near saturation in relatively closed systems where air may not move freely. The resulting situation may need to be described using two flow equations, one for the air phase and one for the liquid phase. The same is true for multiphase air, oil, and water systems in which the fluids are not fully miscible. Flow in such multiphase systems generally require flow equations for each fluid phase involved. Two-phase air-water systems hence could be modeled also using separate equations for air and water. This shows that the standard Richards equation is a simplification of a more complete multiphase (air-water) approach in that the air phase is assumed to have a negligible effect on variably saturated flow, and that the air pressure varies only little in space and time. This assumption appears adequate for most variably saturated flow problems. Similar assumptions, however, are generally not possible when nonaqueous phase liquids (NAPLs) are present. Mathematical descriptions of multiphase flow and transport hence in general require flow equations for each of the fluid phases involved. Assuming applicability of the van Genuchten hydraulic functions and ignoring the presence of residual air and water, the hydraulic conductivity functions for the liquid (wetting) and air phase (non-wetting) phases are given by (e.g., Luckner et al., 1989; Lenhard et al., 2002):

$$K_w(S_e) = K_w S_e^l \left[ 1 - \left( 1 - S_e^{1/m} \right)^m \right]^2 \quad (12a)$$

$$K_a(S_e) = K_a (1 - S_e)^l \left[ 1 - S_e^{1/m} \right]^{2m} \quad (12b)$$

where the subscripts  $w$  and  $a$  refer to the water and air phases, respectively, and  $K_w$  and  $K_a$  are the hydraulic conductivities of the medium to water and air when filled completely with those fluids. A detailed overview of various approaches for measuring and describing the hydraulic properties of multi-fluid systems is given by Lenhard et al. (2002).

### A look ahead

The unsaturated soil hydraulic properties are key factors determining the dynamics and movement of water and its dissolved constituents in the subsurface. Reliable estimates are needed for a broad range of agrophysical applications, including for subsurface contaminant transport studies. A large number of approaches are now available for describing and measuring the hydraulic properties, especially for relatively homogeneous single-porosity soils. This includes direct measurement of discrete  $\theta(h)$ , and  $K(h)$  or  $K(\theta)$  data points and fitting appropriate analytical models to the data, and the use of increasingly sophisticated inverse methods.

Considerable challenges remain in the description and measurement of the hydraulic properties of structured

media (macroporous soils and fractured rock). The hydraulic properties of such media may require special provisions to account for the effects of soil texture and soil structure on the shape of the hydraulic functions near saturation, thus leading to dual- or multi-porosity formulations as indicated by Schaap and van Genuchten (2006) and Jarvis (2008), among others. Estimation of the effective properties of heterogeneous (including layered) field soil profiles also remains an important challenge. Very promising here is the increased integration of hard (directly measured) data and soft (indirectly estimated) information for improved estimation of field- or larger-scale hydraulic properties, including the use of noninvasive geophysical information. New noninvasive technologies with enormous potential range from neutron and X-ray radiography and magnetic resonance imaging at relatively small (laboratory) scales, to electrical resistivity tomography and ground penetrating radar at intermediate (field) scales, to passive microwave remote sensing at regional or larger scales. Challenges remain on how to optimally integrate, assimilate, or otherwise fuse such information with direct laboratory and field hydraulic measurements (Yeh and Šimůnek, 2002; Kowalski et al., 2004; Looms et al., 2008; Ines and Mohanty, 2008), including the optimal and cost-effective use of pedotransfer function and soil texture information, and resultant quantification of uncertainty (Minasny and McBratney, 2002; Wang et al., 2003). These various integrated technologies undoubtedly will further advance in the near future, as well as the use of increasingly refined pore-scale modeling approaches (e.g., Tuller and Or, 2001) at the smaller scales for more precise simulation of the basic physical processes governing the retention and movement of water in unsaturated media.

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## Cross-references

[Bypass Flow in Soil](#)  
[Databases of Soil Physical and Hydraulic Properties](#)  
[Ecohydrology](#)  
[Evapotranspiration](#)  
[Field Water Capacity](#)  
[Hydropedological Processes in Soils](#)  
[Hysteresis in Soil](#)  
[Infiltration in Soils](#)  
[Layered Soils, Water and Solute Transport](#)  
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[Soil Water Management](#)  
[Sorptivity of Soils](#)  
[Surface and Subsurface Waters](#)  
[Tensiometry](#)  
[Water Balance in Terrestrial Ecosystems](#)  
[Water Budget in Soil](#)  
[Wetting and Drying, Effect on Soil Physical Properties](#)

## HYDRODYNAMIC DISPERSION

The tendency of a flowing solution in a porous medium that is permeated with a solution of different composition to disperse, due to the non-uniformity of the flow velocity in the conducting pores. The process is somewhat analogous to diffusion, though it is a consequence of convection.

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## HYDROPEDOLOGICAL PROCESSES IN SOILS

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## Synonyms

Soil hydrophysical processes; Soil physical processes; Soil water processes; Soil water–soil morphology interactions

## Definition

The *hydropedological processes* in the broader sense are all soil processes in which flowing or stagnant water acts as the environment or agent or the vehicle of transport. These processes affect the visible or otherwise discernible morphological features of the soil profile and analogous features on the pedon, polypedon, catena, and soil landscape or soil series scales. These features can be distinguished and categorized according to various pedological classification systems (Lal, 2005) and, vice versa, used to identify and semi-quantify the soil water processes (e.g., Stewart and Howell, 2003) that have produced or affected them. In the narrower sense, only those processes in which water itself (its content, energy status, movement, and balance) is in the focus are regarded as hydropedological processes.

## Introduction

*Pedology* is the branch of soil science dealing with soil genesis, morphology, and classification. In some parts of the world, however, the world *pedology* has been or still is used to denote the whole of soil science. Under these conditions, it was quite natural to name that branch of soil science that deals with soil water (e.g., Stewart and Howell, 2003) and is otherwise referred to, for example, as soil physics, soil water physics, physics of soil water, or soil hydrology as *hydropedology*, notwithstanding its relations (or rather the absence of such relations) to soil

genesis, morphology, and classification. This happened in the fifties of the last century in Czechoslovakia, where the word “hydropedology” was used to denote the discipline of applied soil survey for designing irrigation and drainage systems on agricultural lands (ON 73 6950, 1974; Kutílek et al., 2000). Similar usage may have developed in other countries, too. Recently, the term “hydropedology” was redefined with due regard to both parts of the word, i.e., to the water in the soil and the pedology as defined in the first sentence of this paragraph (Lin, 2003; Lin et al., 2005, 2006a). Hydropedology is thus emerging as a new field, formed from the intertwining branches of soil science, hydrology, and some other closely related disciplines (Lin et al., 2006b, 2008a, b; Lin, 2009). As in hydrogeology, hydroclimatology, and ecohydrology, the emphasis is on connections between hydrology and other spheres of the earth (Wikipedia, 2009), in particular on the pedologic controls on hydrologic processes and properties and hydrologic impacts on soil formation, variability, and functions. Hydropedology emphasizes the *in situ* soils in the context of the landscape (Hydropedology, 2009).

### Hydropedological processes in the soil

The soil and the living or dead vegetation on it transforms the precipitation and snowmelt water into overland flow, infiltration, and evaporation (e.g., Lal, 2005). All these three processes depend not only on the state and properties of the soil on the spot but also on the surface run on and subsurface inflow of water from the upslope parts of the landscape, on the arrangement and properties of soil horizons (e.g., Lal, 2005) (lithologically or pedogenetically generated), and on the boundary conditions at the bottom of the soil (bearing in mind how difficult it is to define any “bottom” of the soil).

The overland flow is generated (in most cases) only locally, due either to the insufficiency of the soil infiltration capacity, the lack of soil permeability when the soil is frozen or covered with ice crust, or to shallow groundwater exfiltrating from the soil in the downslope parts of the landscape, where the soils are often stigmatized by hydromorphism (gleyization, peat horizons, salinization). The overland flow is a vehicle of soil erosion and a carrier of eroded soil particles. The eroded particles are deposited in the places where the overland flow loses its carrying capacity. In this way, the overland flow contributes substantially to the thinning of the upland soils and thickening of the lowland soils and the submerged soils in streams, reservoirs, lakes, and seas.

The shallow subsurface downslope flow often occurs as perched groundwater accumulated on the top of less permeable soil horizons, produced by technogenic compaction or translocation of clay particles (illuviation) or iron and aluminum (lateritization) or iron and organic matter (podsolization) or simply because of the lack of organic matter or the absence of tillage that would render the topsoil more permeable than the subsoil.

The infiltration capacity of the soil is affected, among other factors, by the aptitude of the surface soil to crusting, the roughness of the soil surface and the presence and openness of macropores (Stewart and Howell, 2003) (biopores, cracks, and tillage-induced pores).

The key role in the pedological control of landscape hydrology is played by the retention capacity of the soil profile (e.g., Dingman, 2002). Although it is not exactly true that the soil is capable of retaining all water until its field capacity is exceeded, this rule is nevertheless approximately valid. A nonlinear process, referred to as the “soil moisture accounting,” has to be included in hydrological models in order to turn the infiltration input into the shallow subsurface and deep groundwater runoff output (e.g., Kachroo, 1992). The available water capacity of the soil (the field capacity minus the wilting point) plays also a crucial role in supporting vegetation growth and evapotranspiration.

The field-capacity rule is sometimes vitiated by various types of preferential flow, i.e., a fast gravitationally driven downward movement of water through the spots that are either more permeable or more wettable than the rest of the soils or appear as random manifestations of the hydraulic instability at the wetting front (fingering). This phenomenon is a zone of active research (e.g., Roulier and Schulin, 2008). However, the question of where, when, and to which extent these phenomena occur in different soils and rocks (so that we can predict them) remains largely unanswered.

The hydraulic properties of the soil (Stewart and Howell, 2003), such as the soil moisture retention curve, the saturated hydraulic conductivity, the unsaturated hydraulic conductivity function, the shrinkage curve, the wettability parameters, and many other properties, sometimes easy to quantify but sometimes still resisting to quantification, are mutually correlated and, which is advantageous, are also correlated to other, more easily determinable soil properties such as the particle size distribution, bulk density, and organic matter content (Pachepsky and Rawls, 2004; Pachepsky et al., 2006). It is not incidental that the traditional Czechoslovak “hydropedology” (see above) turned in practice mainly into the particle size distribution analysis of command areas. One task of modern hydropedology would be, in this respect, to reinvestigate the spatial distribution of soil texture classes in conjunction with other soil features, such as the soil depth, soil horizons, the position in the landscape, the degree of hydromorphism, etc.

### Conclusions

The hydropedological processes as a part of soil-water relation processes belong to a new discipline, hydropedology (Lin et al., 2008c). Hydropedology undergoes burgeoning development. Its new topics and subtopics crop up all the time and many existing hot topics can easily accommodate under its wings. In most cases, the acceptance of hydropedological viewpoints is useful and makes the researcher more interdisciplinary and open to new ideas.

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## Cross-references

[Bypass Flow in Soil](#)  
[Ecohydrology](#)  
[Field Water Capacity](#)  
[Hydraulic Properties of Unsaturated Soils](#)  
[Hysteresis in Soil](#)  
[Infiltration in Soils](#)  
[Laminar and Turbulent Flow in Soils](#)

[Overland Flow](#)  
[Pedotransfer Functions](#)  
[Soil Water Flow](#)  
[Water Budget in Soil](#)

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## HYDROPHOBICITY

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See [Soil Hydrophobicity and Hydraulic Fluxes](#)

## Cross-references

[Conditioners, Effect on Soil Physical Properties](#)

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## HYDROPHOBICITY OF SOIL

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## Synonyms

Localized dry spot; Soil water repellency

## Definition

*Hydrophobic* – meaning “water fearing” in Greek.  
*Hydrophobic soils* – repel water, generally resulting in water beaded on the surface.  
*Hydrophobicity* – sometimes refers to a soil–water contact angle  $>0^\circ$ . These soils absorb less water and more slowly than hydrophilic soils.

## Introduction

Hydrophobicity impedes the rate and extent of wetting in many soils. It is caused primarily by organic compounds that either coat soil particles or accumulate as particulate organic matter not associated with soil minerals. Sandy textured soils are more prone to hydrophobicity because their smaller surface area is coated more extensively than soils containing appreciable amounts of clay and silt. The most important effect of hydrophobicity is changes to soil water dynamics. Hydrophobicity causes negative effects through reduced infiltration and water retention, leading to enhanced run-off across the soil surface, preferential flow pathways in the unsaturated zone of the soil, and less plant available water. Many soils that appear to readily take in water have small levels of hydrophobicity.

Reduced wetting rates caused by hydrophobicity may also have a positive impact on soil structural stability. Hydrophobicity can be enhanced by soil drying, heating from fires, soil nutrients, and organic inputs.

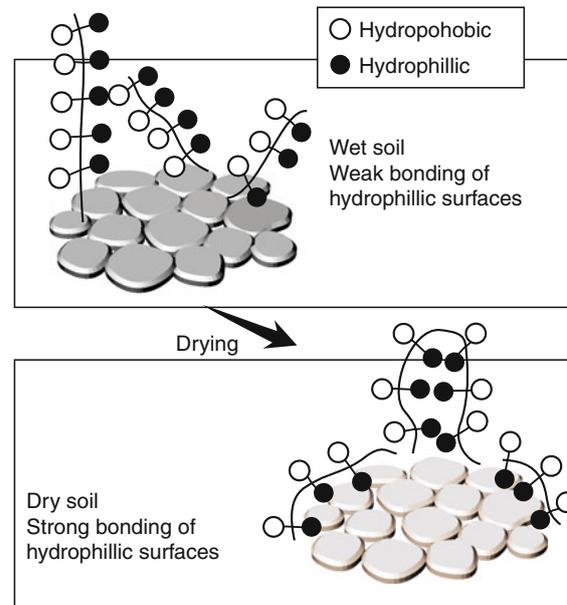
### Geographical occurrence of soil hydrophobicity

Before a surge in research beginning in the 1990s, soil hydrophobicity was generally only associated with semi-arid or coastal soils (DeBano, 2000). The hydrophobicity of over 5 million hectares of agricultural soils in Australia can cause production losses of up to 80% (Blackwell, 2000). It is also a known problem of golf course greens and other sports soils (York and Canaway, 2000). Many coniferous forest soils are extremely prone to soil hydrophobicity (Doerr et al., 2009), particularly following wildfires (Mast and Clow, 2008). Since 1990, greater surveying and the development of more sensitive testing techniques identified soil hydrophobicity as a common property of most soils (Tilman et al., 1989; Doerr et al., 2000). It is now known that temperate soils are affected by soil hydrophobicity, including over 75% of land under pasture and cropping in the Netherlands (Dekker and Ritsema, 1994). Soil hydrophobicity has also been found in subtropical soils (Yao et al., 2009) and can be accentuated by hydrocarbon contamination (Roy et al., 1999). Smaller levels of soil hydrophobicity are found in most soils globally, with soil management (Woche et al., 2005), land use, texture (Doerr et al., 2006), and organic matter (Tilman et al., 1989; Capriel et al., 1995) known to influence the severity. Hydrophobicity tends to increase with decreasing pH, although it has been found in alkaline soils and peats (Doerr et al., 2006).

### Causes of soil hydrophobicity

Long-chain amphiphilic organic compounds produced by a range of biota can induce hydrophobicity in soil (Capriel et al., 1995). These compounds can be highly hydrophilic, but drying causes bonding of hydrophilic (polar) ends of the molecules to each other or soil surfaces, resulting in an exposed hydrophobic (nonpolar) organic surface (Figure 1). Exudates and mycelia produced by fungi have been associated with water repellency in many studies (Bond, 1964; White et al., 2000; Feeney et al., 2006). Plant leaves, root mucilage, algae, and bacterial exudates can also cause soil hydrophobicity (Doerr et al., 2000; Ellerbrock et al., 2009; Martinez-Zavala and Jordan-Lopez, 2009; Hallett et al., 2009). The Lotus effect (Barthlott and Neinhuis, 1997) is an example of extreme hydrophobicity (water drops are not attached to the surface) due to specific combination of hydrophobic waxes and roughness on plant leaves of many plant species.

Although soils may have very different amounts of potentially hydrophobic compounds depending on geography, soil type, and management (Piccolo and Mbagwu, 1999), their concentrations are often poorly related to soil hydrophobicity, particularly if grouped as total organic carbon (Doerr et al., 2000). Severe soil hydrophobicity



**Hydrophobicity of Soil, Figure 1** The polar hydrophilic ends of amphiphilic organic compounds bond to each other and soil particles when dry, resulting in a hydrophobic surface. (Reprinted from Hallett, 2008.)

can result if a few grains have a hydrophobic coating in repacked sands (Steenhuis et al., 2005), but in natural soils the effects could be decreased by cracking of the organic surface during drying, relative humidity impacts, and interactions with other organic compounds (Doerr et al., 2000).

The spacing, packing, and roughness of grains also influence soil hydrophobicity. “Superhydrophobicity,” where water rests on the tips of particles like a bed of nails, has been shown to be a potential process in soils (McHale et al., 2005).

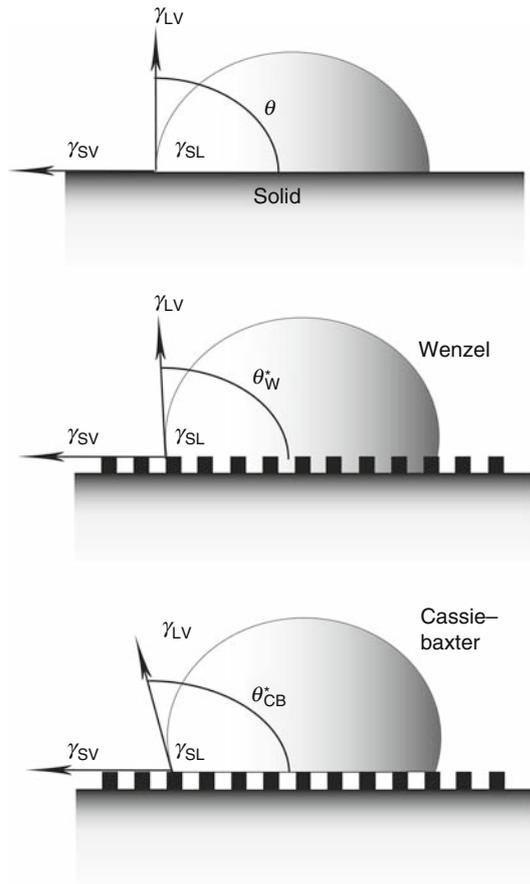
### Physics of water repellency

The contact angle,  $\theta$ , between a drop of water and a solid surface is controlled by the solid–vapor,  $\gamma_{SV}$ , solid–liquid,  $\gamma_{SL}$  and liquid–vapor  $\gamma_{LV}$  interfacial tensions (Figure 2). The Young’s equation

$$\cos \theta = (\gamma_{SV} - \gamma_{SL}) / \gamma_{LV} \quad (1)$$

describes the relation between the contact angle and the interfacial tensions for perfectly flat solid surfaces.

Although contact angles may vary continuously depending on the surface tension of the solid, it is convenient to think in terms of three different wetting situations. Complete wetting, for which the ideal  $\theta$  is zero (perfectly wettable) and the liquid forms a very thin film, partial wetting with  $0^\circ < \theta \leq 90^\circ$  (subcritical water repellency), and non-wetting (severe water repellency) with  $\theta > 90^\circ$ . Roughness of soil particles and pore surfaces can increase  $\theta$  for already hydrophobic soils by either increasing



**Hydrophobicity of Soil, Figure 2** Interaction between a drop of water and solid surface. The solid–vapor,  $\gamma_{sv}$ , solid–liquid,  $\gamma_{sl}$ , and liquid–vapor,  $\gamma_{lv}$  interfacial tensions control contact angle,  $\theta$ . Surface roughness can increase  $\theta$  through Wenzel or Cassie–Baxter processes.

the solid–liquid contact area (Wenzel) or if air within asperities increases the liquid–vapor area at the solid–liquid interface (Cassie–Baxter) (Bachmann and McHale, 2009). These processes can lead to superhydrophobicity and help explain why dry soils are more hydrophobic than wet soils (McHale et al., 2005).

The rate that soil absorbs water is defined by sorptivity,  $S$  and it will be influenced by  $\theta$  as

$$S = S_i \cdot \cos(\theta), \quad (2)$$

where  $S_i$  is the intrinsic sorptivity (Philip, 1957). For a totally non-repellent soil,  $S = S_i$  as  $\cos(0^\circ)$  is 1. Capillarity is influenced by  $\theta$  as

$$z\rho g = \frac{2\gamma \cdot \cos(\theta)}{r}, \quad (3)$$

where  $z$  is capillary rise,  $\gamma$  is the surface tension of water,  $\rho$  is water density,  $g$  is gravity, and  $r$  is the pore radius. A contact angle of  $30$ – $60^\circ$  is not uncommon in soils that

are not recognized as hydrophobic but are to a certain extent water repellent (Woche et al., 2005) and this represents a greater than sixfold drop in sorptivity. Consequently, the observed capillarity rise may be considerably smaller than theoretically expected from Equation 3 with  $\theta = 0$ .

### Measuring water repellency

Numerous approaches exist to measure water repellency in soil (Table 1). The water drop penetration time test (WDPT) is the most commonly used because of its simplicity, suitability for field measurements, and ability to measure the persistence of water repellency over periods of several hours (Dekker et al., 2009). As water repellency is influenced by the hydration status of soil, Dekker et al. (2001) extended the WDPT approach to measure “potential water repellency” using tests on soils equilibrated to different water contents in the laboratory. Usually soil is most water repellent when it is close to its air-dry water content (Dekker and Ritsema, 1994). Severity classes of water repellency can be determined from the WDPT, although some disagreement of critical thresholds exists in the literature. Table 2 provides the most widely accepted WDPT classifications.

The molarity of an ethanol droplet test (MED) is a suitable method for field measurements and indicates how strongly a water drop is repelled by a soil at the time of application (King, 1981). In the MED test, defined surface tensions of water are achieved by varying the molarity with the addition of different amounts of ethanol. The critical surface tension is taken as the minimum ethanol concentration where infiltration occurs in  $<5$  s (Doerr, 1998).

A direct measurement of the soil–water contact angle based on the Sessile Drop method can be achieved using a Goniometer (Bachmann et al., 2000; Diehl and Schaumann, 2007). The capillary rise method (CRM) compares the infiltration rates of water and a liquid (usually hexane) not influenced by hydrophobicity (Bachmann et al., 2003) and is the standard approach used to measure the wetting of powders. A similar concept forms the basis of the intrinsic sorptivity test, where a water repellency index is assessed by comparing the sorptivity of water and ethanol measured with tension infiltrometers (Tilman et al., 1989). The hydrophobicity of individual soil aggregates can be measured by adapting either the CRM (Goebel et al., 2008) or intrinsic sorptivity (Hallett and Young, 1999) methods to assess wetting over smaller surface areas. The approach used to assess the water repellency index can also evaluate the apparent soil–water contact angle (Czachor, 2006). Error in the calculation of apparent soil–water contact angle by CRM or intrinsic sorptivity methods results because of pore roughness and heterogeneity impacts.

### Implications

Soil hydrophobicity is a fundamental physical property of soil that has potentially severe implications to the environment, food security, and land-based industries.

**Hydrophobicity of Soil, Table 1** Major approaches used to assess the hydrophobicity of soil

Test	Approach	Advantages	Disadvantages	Reference
Contact angle "Capillary rise"	Compares wetting rate of water and hexane into a packed column of soil	Physically meaningful Quantifies apparent contact angle	Time consuming Soil is disturbed when packed into columns	Bachmann et al. (2003)
Intrinsic sorptivity or repellency index, <i>R</i>	Compares sorptivity of water and ethanol measured with infiltrometer	Physically meaningful Miniature infiltrometers allow measurements of individual soil aggregates	Interaction between ethanol and soil may influence results	Tilman et al. (1989), Hallett and Young (1999)
Molarity of an ethanol droplet (MED)	Different concentrations of ethanol in water applied as drops to soil surface Critical minimum "molarity" where rapid infiltration occurs	Quick and easy (10 s per test)	Physical meaning requires greater investigation Surface roughness influences results	King (1981), Dekker et al. (2001), Roy and McGill (2002)
Sessile drop	Optical measure of contact angle of water drop on soil surface using Goniometer or light microscope	Measures contact angle directly	Affected by surface roughness Difficult to measure on wettable soils	Bachmann et al. (2000), Diehl and Schaumann (2007)
Water drop penetration time (WDPT)	Infiltration time of a drop of water placed on the surface of soil	Easily measures the persistence of hydrophobicity	Affected by pore structure Takes considerable time in repellent soil Not sensitive enough for low levels of hydrophobicity	Dekker et al. (2009)
Wilhelmy plate method (WPM)		Measures both advancing and receding contact angles	Uses a disturbed sample. Impact of adhesive and glass slide	Woche et al. (2005)

**Hydrophobicity of Soil, Table 2** Classes of water repellency defined by Dekker et al. (2001) for the water drop penetration time (WDPT) test

Class	Severity	WDPT
0	Wettable, non-repellent	<5 s
1	Slightly water repellent	5–60 s
2	Strongly water repellent	60–600 s
3	Severely water repellent	600–3,600 s
4	Extremely water repellent	1–3 h
5		3–6 h
6		>6 h

The decreased rate of water infiltration and retention caused by hydrophobicity results in greater overland flow, less water retention, and the development of preferential flow paths and patchy dry spots in soil. Conventional soil physics approaches to describe water transport and retention require extensions to be effective in soils exhibiting even small levels of hydrophobicity (Deurer and Bachmann, 2007).

On golf courses, soil hydrophobicity is prominent and exacerbated by nutrient inputs and the small surface area of sand grains used to form putting greens (York and Canaway, 2000). Drought-stressed grass develops over hydrophobic soils as plant available water is reduced severely. In severely water stressed countries, the long-term irrigation of soil with treated effluent (waste water)

can present a serious challenge if hydrophobicity causes poor delivery and retention of water in the root zone (Wallach et al., 2005; Graber et al., 2006; Vogeler, 2009).

Increased overland flow due to hydrophobicity accentuates soil erosion (Scott and Van Wyk, 1990; Shakesby et al., 2000; Benavides-Solorio and MacDonald, 2005), particularly following forest fires (Osborn et al., 1964). The impact follows seasonal shifts in hydrophobicity, with the impacts greatest during the summer (Witter et al., 1991; Jungerius and ten Harkel, 1994). Raindrops on hydrophobic soils produce fewer, slow-moving ejection droplets compared to wettable soils but remove more sediment (Terry and Shakesby, 1993). With successive drops, the surface of hydrophobic soils remain dry and noncohesive, leading to displacement by rain splash despite the overlying film of water (Doerr et al., 2003; Leighton-Boyce et al., 2007).

There is anecdotal evidence that heavy precipitation following dry periods can lead to flooding due to soil hydrophobicity. With increasing drought and severe weather predicted with climate change, this could have severe implications, particularly if predicted increases in the frequency and severity of soil wetting and drying occur.

Not all implications of soil hydrophobicity are deleterious to the environment or food production. Slower wetting rates of soil caused by hydrophobicity can result in increased soil aggregate stability (Goebel et al., 2005). Evaporation is also decreased by hydrophobic surface soil

(Shokri et al., 2009) and this mechanism is used by microbiotic crusts to conserve water in extremely arid environments (Issa et al., 2009).

### Amelioration

Physical, chemical, and biological approaches have been developed to combat problems associated with hydrophobic soils. As soil hydrophobicity is pH dependent, lime application can reduce acidity of soils and is a very common amelioration strategy. Wetting agents are also in widespread use (Oostindie et al., 2008), particularly on amenity soils (Cisar et al., 2000) but also increasingly on agricultural land (Miyamoto, 1985). They can improve water distribution and infiltration rates by either acting as surfactants that decrease the surface tension of water or by altering the contact angle of soil surfaces (Kostka, 2000). A common approach on hydrophobic agricultural soils is the addition of clay to cover hydrophobic surfaces and make them hydrophilic (Blackwell, 2000). Kaolinitic clays are the most effective in reducing repellency (Ma'shum et al., 1989; Ward and Oades, 1993; McKissock et al., 2002; Dlapa et al., 2004), but relatively large quantities of clay are required to achieve the desired effect ( $100 \text{ t ha}^{-1}$ ) (Blackwell, 1993), so the approach is only economical if the clays occur naturally on site. Furrows can also help combat the impact of hydrophobicity by harvesting water and diverting it to the root zone (Blackwell, 2000).

Intensive cultivation decreases soil hydrophobicity as organic coatings are abraded and new soil surfaces are exposed. However, the shift in microbial dynamics and carbon mineralization that ensues can lead to hydrophobicity developing again (Feeney et al., 2006). Zero or reduced tillage, on the other hand, can decrease soil hydrophobicity by maintaining greater soil moisture (Blackwell, 2000) and potentially by altering the functional capacity of microbes to degrade hydrophobic compounds (Roper, 2005). Wax degrading bacteria have been isolated that have been demonstrated to reduce soil hydrophobicity (Roper, 2004).

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## HYPOBARIC STORAGE

Hypobaric storage involves the cold storage of fruit under partial vacuum. Typical conditions include pressures as low as 80 and 40 millimetres of mercury and temperatures of 5° C (40° F). Hypobaric conditions reduce ethylene production and respiration rates; the result is an extraordinarily high-quality fruit even after months.

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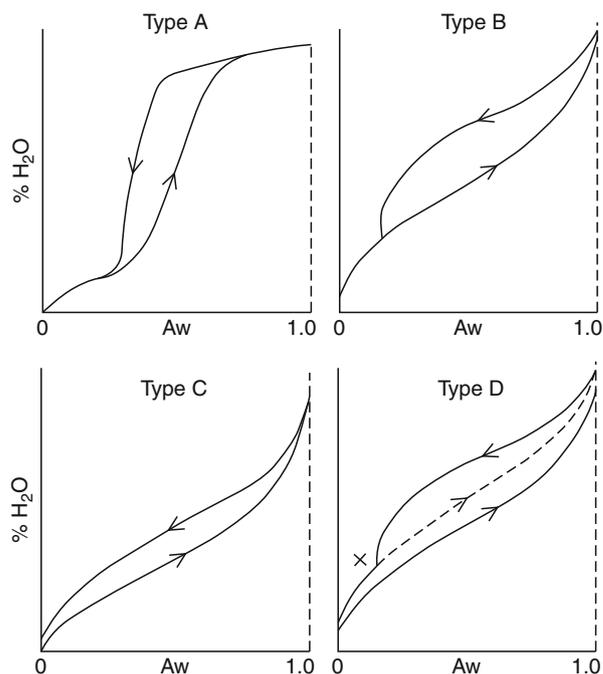
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## HYSTERESIS IN FOODS

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Hysteresis in foods is the phenomenon by which at constant water activity ( $A_w$ ) and temperature, a food adsorbs a smaller amount of water during adsorption than during a subsequent desorption process. Previous hypotheses to explain the phenomenon (Zsigmondy, 1911; Cohan, 1938, 1944; Everett, 1967; McBain, 1935; Kraemer, 1930) have been based on capillary condensation but the phenomenon is exhibited in foods and other substances believed to have negligible capillaries or pores. Current explanation (Caurie, 2007) states that sites adsorb moisture appropriate to their surface energies. During adsorption, micro-cracks and fissures form in the food to expose additional sites. Exposed sites unable to adsorb moisture on the way up to higher water activities ( $A_w$ ) because of inappropriate surface energies adsorb additional moisture on return to lower  $A_w$ s at appropriate  $A_w$  and surface energies to exhibit a hysteresis loop.



Hysteresis in Foods, Figure 1 Types of hysteresis.

The equation derived to fit the above interpretation of hysteresis has the form (Caurie, 2007)

$$m\left(\frac{1}{a} - 1\right) = K \exp\left(B\frac{Q}{T}\right), \quad (1)$$

where  $m$  = percent moisture content at water activity  $a$ ;  $Q$  = surface energy (keal/mole);  $T$  = absolute temperature;  $K$ ,  $B$  = constants.

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## HYSTERESIS IN SOIL

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### Synonyms

Soil water hysteresis

### Definition

Hysteresis in soil is defined as the difference in the relationship between the water content of the soil and the corresponding water potential obtained under wetting and drying process.

The relationship between soil water content and soil water potential is called soil water retention curve (SWRC). This dependency manifests itself through hysteresis. It was shown by Haines in 1930 (Haines, 1930). This means that water content in the drying (or drainage) branch of water potential – water content relationship – is larger than water content in the wetting branch for the same value of water potential. For hygroscopic water, this effect is due to differences of water content at increasing and decreasing vapor tension in soil. During the increase of vapor tension, the water content in soil is lower than during the vapor tension decrease. For capillary water, the hysteresis phenomena result from pore shape irregularity. Irregular soil capillary is characterized by volume  $V$  and minimal  $r$  and maximal  $R$  radiuses. Empty capillary is filled with water at under pressure corresponding to radius  $R$ . After filling with water, the meniscus is created, corresponding to radius  $r$  and the same capillary can be emptying at much higher water under pressure. The hysteresis region is called hysteresis loop. The wetting and

drying curves can be of the first or higher orders depending on actual soil water potential at which the wetting or drying process is started. Numerous models describing soil water hysteresis were developed. These models can be categorized into two main groups: the conceptual models and empirical models. The conceptual models are basically based on the domain theory. The independent domain theory of soil water hysteresis assumes that each soil water domain wets and dries at the characteristic water potentials irrespective of neighboring domains. This theory has been developed by Néel (1942, 1943). The modification of this theory takes into account interaction between particular domains and in literature is referred as dependent domain theory (Poulovassilis and Childs, 1971; Topp, 1971; Mualem and Dagan, 1975). Empirical models are mainly related to the analysis of the shape and properties of water retention curves.

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Encyclopedia of Agrophysics

Gliński, J.; Horabik, J.; Lipiec, J. (Eds.)

2011, XLVI, 1028 p. 421 illus., 168 illus. in color.,

Hardcover

ISBN: 978-90-481-3584-4